Drivers of continental margin growth.
Examples from the Quaternary Adriatic Basin.

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Preface

The growth of continental margins typically results in the development of large-scale sedimentary units with sigmoidal-shaped profiles, termed clinothems. Clinothems occur at a variety of spatial and temporal scales, are recognized from shelfal to basin regions and represent one of the fundamental building blocks of prograding successions. This Ph.D. thesis is based on four datasets from different case histories along an hypothetical sediment routing system of the Adriatic continental margin. The four case histories allow to investigate how the sediment is transferred from river to continental margin and to document the depositional architecture of a set of prograding successions developed in three different environments, from coastal-deltaic to deep marine settings. The datasets available consist of orthophotos, multibeam bathymetry, Chirp sonar and multichannel profiles, sediment cores and well logs data. The main goal of this Ph.D. thesis is to investigate how the interaction of allogenic and autogenic processes impacts on river morphodynamics and drives clinothems development and growth, and ultimately the formation of modern continental margins. The first case history is a portion of the Po River upstream the Isola Serafini dam. This area may be considered as a natural laboratory where investigate and quantify the impact of the backwater effect on river morphodynamics. The second case history is the Mid Adriatic Deep during the Last Glacial Maximum, where in a river-dominated system developed a 350 m thick succession in a very short-time window. This area allows to investigate the internal architecture, geometric relation and facies distribution of a lowstand delta with high chronological resolution. The third case history, offshore the Gargano Promontory, offers the possibility to investigate how genetically-related coastal and subaqueous progradations, i.e. a compound delta, may develop at sub-millennial time scale. The fourth case history, the southern Adriatic Basin, gives opportunity to investigate the impact of the oceanographic regime on sediment transport, where, far from direct sediment feeding sources, lateral advection and current deposition became the dominant mechanism of margin progradation.

Working on the Adriatic Quaternary succession it was possible to document the effect of a paramount control on margin feed and progradation, to discriminate short-lived phases of extremely rapid deposition during the margin construction, and to document the importance of oceanographic processes on margin growth.
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“The same parts of the earth are not always moist or dry, but they change according as rivers come into existence and dry up. And so the relation of land to sea changes too and a place does not always remain land or sea throughout all time, but where there was dry land there comes to be sea, and where there is now sea, there one day comes to be dry land. But we must suppose these changes to follow some order and cycle. The principle and cause of these changes is that the interior of the earth grows and decays, like the bodies of plants and animals...”

(From Meteorology, Aristotele 340 BC ca.)
1) Introduction

1.1) Continental margins

The boundary between continents and the oceans occurs in a physiographic region known as continental margin. Continental margins extend from coastal plains to deep-sea basins and encompass shorelines, shallow shelves and steep slopes (Fig. 1.1). Passive continental margins attain the largest sediment accumulation rates on Earth, which results in thick successions that have the potential for resolving signals of change in accommodation, oscillation in sediment flux and variation in oceanographic regime paced over a range of time-scales. For these reasons stratigraphic successions of continental margins represent extraordinary archives from which extract the record of continental-margin processes, climate and paleoceanography (e.g. Pratson et al., 2007).

Figure 1.1: map of the total sediment thickness of the World’s oceans and Marginal Seas (modified from https://www.ngdc.noaa.gov/mgg/sedthick/sedthick.html). Bottom left: the passive continental margin profiles with coastal plain, broad continental shelf, and continental rise (from Brink et al., 1992).
Rivers deliver some 84% of the total sediment load that reaches the oceans (Milliman and Meade, 1983) and are the dominant means by which clastic sediments are transported to continental margins. Rivers have been active in delivering sediments to continental margins since their inception, but individual rivers are ephemeral on geological timescales. Tectonics, glaciations and stream pirating among other processes alter river courses or have created new rivers from old ones. So, while fluvial input to margins may be considered continuous, the cast of rivers responsible for this input is subject to change (Bridge, 2003). Large individual rivers, such as the Amazon, Indus, Yangtze and Mississippi, often act as a single point source of sediment to a continental margin. Where multiple rivers feed a margin, they can combine to form a single line source (Jaeger and Nittrouer, 2000; Walsch et al., 2004; Cattaneo et al., 2007; Liu et al., 2009; Patruno et al., 2015). Sediment emanates from the mouth of each river, but on the shelf these contributions are smeared together by waves and currents, making it very difficult to define where the input sediment comes from.

Figure 1.2: stratigraphic stacking patterns associated with changing rates of coastal accommodation creation (δA) and sediment fill (δS) (modified from Hunt and Tuker, 1992; Neal and Abreu, 2009).
The integration of seismic and sequence stratigraphy is a powerful method in revealing the main factors that govern the outbuilding of continental margins (Vail et al., 1977; Vail, 1987; Van Wagoner, 1988). Sequence stratigraphy is focused on analyzing changes in geometric characters and seismic facies of strata packages and on the identification of key bounding surfaces to determine the chronological order of basin fill and erosional events (Catuneanu et al., 2009). Stratal stacking patterns respond to the interplay of changes in the rate of sediment supply and base level (through tectonic subsidence, isostasy, sediment compaction and eustasy), and reflect combinations of depositional trends (Steckler al., 1999). Each stratal stacking pattern defines a particular genetic type of deposit defined as transgressive, normal regressive and forced regressive (Hunt and Tucker, 1992); alternatively, strata packages can be defined as sets of retrogradation, progradation to aggradation, aggradation to progradation to degradation deposit (Fig. 1.2; Neal and Abreu, 2009).

The relationship between stratal stacking pattern and cyclic change in base level is a fundamental theme of sequence stratigraphy; the concept of “base level” delineates a dynamic surface of balance between erosion and deposition (Neal and Abreu, 2009), and is strictly related to the concept of ‘accommodation’, i.e. the amount of space available for sediment accumulation (Jervey, 1988; Neal and Abreu, 2009): a rise in base level creates accommodation, whereas a fall in base level destroys accommodation. Base level is commonly approximated to relative sea level (e.g., Jervey, 1988), although it can lie below sea level depending on the energy of the marine system, especially waves and currents (Pirmez et al., 1998; Swenson et al., 2005). When base level approximates sea level, the concept of ‘base-level change’ becomes equivalent to the concept of ‘sea-level change’ (Posamentier et al., 1988; Swenson et al., 2005) and may become a powerful tool for reconstructing the paleo sea-level oscillations up to a global scale (Haq et al., 1987).

After the early to middle Pleistocene transition, high-amplitude (100 m) sea level fluctuations, among other factors, played a crucial role in continental margins outbuilding (Vail et al., 1977; Haq et al., 1987; Trincardi and Coreggiari, 2000; Schlager, 2004; Rabineau et al., 2005; Ridente et al., 2009; Schattner et al., 2010; Lobo and Ridente, 2014). These fluctuations have many impacts, in
particular forcing the migration of the shoreline hundreds of kilometers basinward until reaching the
shelf edge during periods of sea level lowstand, promoting coarse-sediment delivery to the deep sea
(Covault and Graham, 2010). Conversely, sea level rise causes landward migration of the shoreline
(defined as the region from the low-tide shoreline to ca. 10 m water depth), leading to the development
of backstepping coastal lithosomes (e.g. Cattaneo and Steel, 2003). The slowing of the rate of sea
level rise at the end of each cycle allows most river systems to fill their estuaries, and is followed by
a broadening of the sediment dispersal system onto the continental shelf to form deposits with a
prominent morphological expression. During periods of sea level highstand a great amount of sand
delivered by rivers is trapped on the inner shelf forming subaerial and coastal deposits while finer
sediment fractions are transported farther seaward forming subaqueous depocenters (e.g. Michels et
al., 1998; Goodbred and Kuehl, 2000; Liu et al., 2009; Korus and Fielding, 2015). Modern deltas and
subaqueous clinoforms, originated during the last 5-6 thousands of years (Stanley and Warne, 1994)
are observed in different tectonic settings: on passive margins, such as the Amazon deltaic system
(Nittouer et al., 1986); on active-margins, such as the margin where the Ganges-Brahmaputra deltaic
system is located (Goodbred et al., 2003); and broad epicontinental shelves such as the Yellow River
deltaic system (Alexander et al., 1991). Moreover, these deposits usually exhibit an asymmetrical
thickness distribution with respect to their feeding system(s) and reflect the influence or dominance
of lateral sediment dispersion through advection (McCave, 1972; Nittouer et al., 1996; Bhattacharya
and Giosan, 2003; Cattaneo et al., 2007).

Eustatic oscillations have occurred in association with climatic fluctuations throughout the
Quaternary (Shackleton et al., 1984; Mix and Ruddiman, 1985; Chapman and Shackleton, 1998; Mix
et al., 2001), and in turn affected changes in sediment supply to the basin (Mulder et al., 2013),
variations in oceanographic regime (Toucanne et al., 2007), and basin interconnectivity (Hernández-
Molina et al., 2014). Among all the sedimentary bodies present on continental margins, bottom-
current-deposits (i.e. contourites) represent excellent archives for paleoclimatic and
paleoceanographic reconstructions because of their fairly continuous sedimentation rate and their
high potential in offering high-resolution chronostratigraphy (Faugères et al., 1999; Rebesco et al., 2014 for a review).

Recent advances in geophysical-survey and techniques, especially the use of very high-resolution seismic and 3D multibeam bathymetry, provided the opportunity to analyze the stratigraphic succession of continental margins with higher detail and improved accuracy, both in shallow and deep waters. These advances in knowledge led to a better understanding of the main processes that control the formation of stratigraphic successions and of the evolution of continental margins as a whole.

The overarching aim of this thesis is to contribute to the understanding of the main processes that, by interacting with each other, governed the dispersion of sediments and the final architectural motif of the Quaternary sedimentary succession of the Adriatic continental margin. A wealth of geophysical and core data from the region were analyzed in a multi-methodological approach in order to:

- understand the hydrodynamics of a modern river with the ultimate aim to provide new insights on the impact of backwater dynamics on the construction of river deltas (chapter 3);
- evaluate the main factors that generated changes in accommodation on millennial time scales, in a very high sediment supply deltaic system during a lowstand sea level (chapter 4);
- understand the role of lateral advection in delta construction during the post glacial maximum sea level rise (chapter 5);
- unravel the role of changing bottom current intensity and pathways from the Pliocene to late Holocene on the construction of a continental margin (chapter 6).
1.2) High-frequency cyclicity in the sedimentary record of continental margins

It is widely accepted that climatic fluctuations reflect the variations in the astronomic parameters that define the Earth’s motion around the Sun (eccentricity, obliquity and precession; Milankovitch, 1930), as expressed in the modulation of solar radiation with periodicities of ca. 100, 40 and 20 kyr, respectively (Chappell and Shackleton, 1986; Schwarzacher, 2000). Since their first detection in the marine geological record (Emiliani, 1995), these Milankovitch cycles have been shown to drive sea level variations as expressed in Marine Isotopic Stage (MIS), and Substages, of Quaternary records (e.g. Hays et al., 1976; Siddal et al., 2003). The Milankovitch cyclicity can be identified in a variety of sedimentary records, as documented during the last two decades (Mitchum and Van Wagoner, 1991; Hernández-Molina et al., 2000; Jouet et al., 2006; Ridente et al., 2009; Zhu et al., 2012; Lobo and Ridente, 2014 among others), and on continental margins, as major high-frequency base level changes. Base level cycles were predominantly paced by 40 kyr interval during the Pliocene and Lower Pleistocene (Carter et al., 1998), while sea level and hence sediment deposition has been strongly controlled by 100 and 20 kyr cyclicity after the early to middle Pleistocene transition (Fig. 1.3; Maslin and Ridgwell, 1995).

Most of the sedimentary successions on modern continental margins show a dominant influence of 100 kyr glacial-interglacial cycles, including well-studied Mediterranean areas (e.g. Cattaneo and Trincardi, 1999; Amorosi and Colalongo, 2005; Rabineau et al., 2005). Each sequence is bounded by shelf-wide unconformities recording subaerial erosion (e.g. Hernandez-Molina et al., 1994; Trincardi and Correggiari, 2000; Tesson et al., 1998; Ridente and Trincardi, 2002; Bernè et al., 2007).
Recently, sub-Milankovich climate and sea level fluctuations have been documented in the Gulf of Lions, where short-lived prograding sequences could be linked to 5-10 kyr cycles matching the occurrence of Heinrich events (Bassetti et al. 2008; Sierro et al., 2009); while changes in the inflow, ventilation and vertical fluctuation punctuated the Levantine Intermediate Water and variations in temperature impacted the deep-water formation along the Western Mediterranean margin, matching the timing of Dansgaard-Oeschger cycles (Cacho et al., 2006; Toucanne et al., 2007; Minto’o et al., 2014).
Throughout the Holocene climatic fluctuations known as Bond events occurring every ca. 1.5-2 kyr were identified primarily from fluctuations in ice-rafted debris by studying petrologic traces of drift ice in the North Atlantic; these Bond events may be viewed as the interglacial counterpart of the glacial Dansgaard-Oeschger events (Bond et al., 1997). A similar cyclicity was documented to affect discharge rates of the Po delta in the Adriatic Sea through the Holocene (Piva et al., 2008).

A peculiar example of abrupt climate change occurred within late Pleistocene during the cold spell of the Younger Dryas (Fairbanks, 1989). The Younger Dryas stadial, between 12.8-11.6 ka BP (Severinghaus et al., 1998), was a major cooling event linked to an abrupt interruption of the thermohaline oceanographic circulation in the North Atlantic (Berger, 1990) that marked the transition between two steps of deglacial warming characterized by pulses of meltwater input (Bard et al., 1990). The meltwater pulse 1A (MWP-1A), at about 14.08-13.61 ka BP, centered with the Older Dryas, and the MWP-1B, centered at 11.4–11.1 BP during the last phases of the deglaciation (Berger, 1990; Camoin et al., 2012). The Younger Dryas event was recorded in the stratigraphy of Mediterranean margins by a set of shelf-phase deltas and subaqueous progradations, offshore, accompanied by the re-incision of channel-belt deposits by rivers, on land (Trincardi et al., 1996; Cattaneo and Trincardi, 1999; Labaune et al., 2005; Berné et al., 2007; Hernández-Molina et al., 1994; Amorosi and Milli, 2001; Maselli et al., 2011; Sømme et al., 2011).

The identification of bounding surfaces relating to Milankovitch- and Sub-Milankovitch-type depositional units is of great importance in the investigation of prograding sequences since, as mentioned above, these may reflect eustatic oscillations, sudden changes in sediment supply, and shifts in the oceanographic regime. The combination of these fluctuating environmental factors becomes especially important in river-dominated systems (chapter 4), supply- and advection-dominated deltaic systems (chapter 5) and bottom-current-dominated systems (chapter 6), where climatically-induced perturbations may be recorded by changes in sedimentary stacking pattern.
1.3) Adriatic geological setting

1.3.1) Physiographic setting of the Adriatic basin

The Adriatic Sea is an epicontinental basin elongated about 800 km long in NW-SE direction and about 200 km wide, with a present day surface area of ca. 138,600 km² (Cushman-Roisin, 2001) that represents the northernmost part of the Mediterranean Sea extending as far as 45°47' North. The northern Adriatic Sea is characterized by a low-gradient profile (ca. 0.02°) that reaches its maximum depth of about 252 m in the Mid Adriatic Deep (MAD), a remnant slope basin aligned in a SW-NE (Fig. 1.4). South of the MAD, localized bathymetric irregularities, such as the Tremiti Islands, the Gargano Promontory, the Gallignani Ridge and the Pelagosa Sill, are the morphological expression of recent tectonic activity (Fig. 1.4). The southern Adriatic Sea, beyond the Pelagosa Sill, reaches depths of ca. 1200 m and is flanked by a steep slope and a broad shelf, about 70-80 km south of the Gargano Promontory (Fig. 1.4).

Figure 1.4: Left: Digital elevation model of the Adriatic Sea with main structural elements. MAD stands for Mid Adriatic Deep. Right: slope angle map of the Adriatic Sea floor shallower than 300 m. The green dashed line indicates the offlap break of the Late Holocene deposits (Cattaneo et al., 2007). Note that the shelf is steeper in the study area than in the northern Adriatic. The digital elevation model of the land surface is derived from SRTM 90 m Digital Elevation Data (http://srtm.csi.cgiar.org). Modified after Maselli et al., 2011.
The steep and irregular western slope of the southern Adriatic Sea is characterized by an uneven morphology related to several structural highs (e.g. the Gondola anticline and the Dauno sea mount), and large incisions, as the Bari Canyon (Fig. 1.4). The southern Adriatic Sea is connected with the Ionian Sea through the Otranto Strait, a relatively narrow and deep sill (72 km wide and 780 m deep), which plays a fundamental role in determining the present-days circulation pattern of the whole Mediterranean Sea (Artegiani et al., 1997a,b).
1.3.2) Tectonic setting of the Adriatic basin

The Adriatic basin lies on a continental crust (Adria), located between the East-verging Apennine and the West-verging Dinarides (Ori et al., 2009; Royden et al., 1987). The structural setting and sedimentary architecture of the Adriatic basin reflect a long-term evolution from a passive margin during the Mesozoic to a foreland domain during the Cenozoic (D’Argenio and Horvath, 1984). The western Adriatic margin of the Adriatic basin represents the Oligo-Miocene to early Pleistocene Apennine foreland basin (Fig. 1.5; Ricci Lucchi, 1986; Argnani et al., 1993, 2001, 2009; Doglioni et al., 1994, 1996).

The emerged sectors of the Gallignani Ridge, the Gargano Promontory and the Apulia ridge correspond to the flexural bulge resulting from the subduction activity connected to the Apennine and Dinaric-Ellenic belts. (Argnani et al., 1993, de Alteriis, 1995; Bertotti et al., 1999). North of the Gargano Promontory, the central Adriatic shelf area is bounded by NW-SE trending Gallignani and Pelagosa ridges, which correspond to regional folds along this peripheral bulge (de Alteriis, 1995).

The tectonic domain provides evidence of marked structural differentiation from North to South of the Gargano, likely related to deep crustal discontinuities and variations in lithosphere thickness of the westward dipping Adriatic plate (Doglioni et al., 1994; de Alteriis, 1995). Transpressive structures occur in the North within the NE-SW oriented Tremiti-Pianosa High (Argnani et al., 1993;
de Alteriis, 1995) and along the Gargano Deformation Belt, extending offshore in the W-E trending Gondola fault (Ridente et al., 2008).

The tectonic evolution of the central Adriatic Sea since the Oligocene shows that the highest subsidence values are confined landward, toward the Apennine foredeep, as highlighted by the flexure of the Messinian datum (e.g. Scrocca, 2006) and the average subsidence rates of 0.3 mm/yr calculated from the PRAD 1-2 borehole placed in the central Adriatic Sea (Maselli et al., 2010). Since the middle Pleistocene, the units infilling the foredeep basin have changed from a dominant turbidite fill into a progradational margin wedge that records the Milankovich glacial-eustatic cyclicity (Ridente et al., 2009; Ghielmi et al., 2010).

South of the Gargano Promontory the South Adriatic Margin corresponds to the present days Dinarides Foredeep and the Bradanic trough, on the western side of the Apulia ridge, corresponds to the Apennine foredeep (Argnani et al., 1996). Combined seismic and outcrop data suggest that, during the Pliocene, a large and subsiding foredeep basin encompassed the areas North and South of the Gargano Promontory (Torre et al., 1988). In the late Pliocene, the growth of structural highs mainly related to the establishment of the Tremiti transfer zone, separated this foredeep into two distinct basins: the Adriatic foredeep basin and the Bradanic foredeep basin, respectively North and South of the Gargano Promontory (Capuano et al., 1996). These two sectors were then further decoupled with the growth of the Apulia Ridge, during the middle-late Pleistocene (Doglioni et al., 1994).

In the southern Adriatic Sea, deformation since the Pliocene was likely related to the re-activation and inversion of inherited Mesozoic extensional faults (Argnani et al., 1993). These faults form the Gargano Deformation Belt, which develops both on land with the Monte S. Angeolo-Mattinata fault, and offshore where the W-E-trending deformation zone related to the Gondola fault and anticline are the most evident features (Billi and Salvini, 2000; Ridente et al., 2008).
1.3.3) Oceanographic setting of the Adriatic basin

The Adriatic Sea is a micro-tidal epicontinental sea dominated by a cyclonic circulation driven by thermohaline processes (Artegiani et al., 1997a, b; Poulain, 2001). Three water masses define the general oceanographic setting (Fig. 1.6; Paschini et al. 1993): (1) a surficial temperature-mixed layer (0-30 m); (2) a Levantine Intermediate Water (LIW); (3) and the Northern Adriatic Deep Water (NAdDW).

The cyclonic circulation of surficial waters, with the upper 10 m of less saline and cooler waters of coastal origin, mainly formed by the Po River runoff, forces the fresh waters to flow along the western side of the basin. Fluvial sediment sources are located almost exclusively along the north and western side of the Adriatic basin, with a combined modern delivery of 51.7 x 10^6 t yr-1 of mean suspended load with contributions of 3 x 10^6 t y-1 from eastern Alpine rivers, 15 x 10^6 t y-1 from the Po river, 32.2 x 10^6 t y-1 from the eastern Apennine rivers and 1.5 x 10^6 t y-1 from rivers south of the Gargano promontory (Frignani et al., 1992; Milliman and Syvitski, 1992; Cattaneo et al., 2003). In this shallow part of the basin, hypoxic events occur, typically, between September and November and reflect low current velocities at the gyre centre, enhanced water stratification, which reduces vertical mixing, and high turbidity which prevents light penetration. Currents are stronger away from the gyre centre, resulting in a prevailing flow to the SE along the Italian coast, a flow that appears consistent with the overall shore-parallel thickness of the basin.
distribution of the late-Holocene clinoform (Cattaneo et al., 2003 and 2007). Sediment transport is enhanced in the northern Adriatic by the northeasterly Bora wind that enhances waves and currents especially in winter, with an average southward transport along the shelf and reduced across-shelf transport (Lee et al., 2005).

The LIW, a denser and salty water mass (29.0 kg/m$^3$) that forms in the Levantine Basin through evaporation during the summer and cooling during winter (Lascaratos et al., 1999), enters in the Adriatic Sea through the Otranto Strait and flows with a cyclonic path in a water depth range between 200 and 700 m (Fig. 1.6; the so-called South Adriatic Gyre: e.g. Arbegiani et al., 1997a, b; Mantziafou and Lascaratos, 2008).

The North Adriatic Dense Water (NAdDW), the densest water in the whole Mediterranean with densities up to 1.030 kg/m$^3$ and mean temperatures of ~ 11 °C during extreme events (Vilibič, 2003), is generated on the broad and shallow (< 40-m-deep) North Adriatic shelf (Fig. 1.6) through intense cooling and evaporation during January and February and is largely affected by severe winter outbreaks of the cold and dry northeast Bora wind (Ivančan-Picek and Tutiš, 1996; Grubišič, 2004), and by river runoff. Due to its spatial heterogeneity, the Bora wind produces a coupled cyclonic–anticyclonic gyre system in the North Adriatic (Hendershott and Rizzoli, 1976; Arbegiani et al., 1997a, b), thus increasing vertical mixing in the area, especially in the centre of the cyclonic gyre. The newly formed NAdDW flows southward as different water branches along the Italian coast, towards the MAD and the South Adriatic Margin (SAM; Arbegiani and Salusti, 1987). On its routing, the NAdDW may sink and flow as a bottom-trapped current underneath the older water masses present in the MAD: lifting and pushing the older water masses through Pelagosa Sill, and promoting water exchange among the basin and the migration of small-scale sediment drifts (Marini et al., 2015). Typically, the NAdDW reaches the South Adriatic slope two months after its generation (Vilibič and Orlič, 2001), strongly interacts with the slope topography by enhanced turbulent mixing, and finally sinks to the bottom of the SAM (Trincardi et al., 2007; Benetazzo et al., 2014). Through this process the NAdDW cascades obliquely across the south Adriatic slope along the steepest sector reaching
below the depth range impacted by the contour parallel LIW (Trincardi et al., 2007; Canals et al., 2009). On the other side, due to its dynamical properties (buoyancy and kinetic energy), a portion of the NAdDW remains trapped on the shelf, flowing as a contour-parallel bottom current over several hundreds of kilometers (Fig. 1.6; Benetazzo et al., 2014; Bonaldo et al., 2015).

Both the contour-parallel LIW and the NAdDW interact with the seafloor morphology along preferred paths, leading to the formation of depositional and erosional features in water depths between ca. 100 and 1.200 m (Fig. 1.6; Verdicchio et al., 2007; Foglini et al., 2015). Recent analysis of erosional and depositional features highlighted how the process of bottom-hugging currents concur to a thorough “restyling” of the submarine landscape markedly differentiated morphologies and sediment distribution (Foglini et al., 2015). These processes were active, or become re-activated, after the end of the last glacial maximum (Verdicchio et al., 2007), likely in response to the sea level rise that flooded the Adriatic shelf thereby permitting the formation of dense shelf waters by rapid winter cooling of an extensive shallow-water region (less than 50 m deep).
2) **Overview of own research**

The purpose of this PhD project is to understand the main processes that governed the evolution of the Adriatic basin during the Quaternary and to define allogenic and autogenic signals preserved in the stratigraphic record. To achieve this goal, four case histories were selected along the Adriatic basin (Fig. 2.1) where relative impact of sediment flux and basin dynamic are markedly differentiated: 1- (northern Adriatic Po plane): the effect of backwater on the morphodynamic evolution of a river (chapter 3); 2- central Adriatic margin (Adriatic shelf and Mid Adriatic Deep): river-dominated system with very high sediment supply, where direct supply from the catchments is the major controlling process (chapter 4); 3- central Adriatic margin (offshore the Gargano Promontory): a supply-and advection-dominated system, where fine grain sediments are transported far from the main feeding point (chapter 5); 4- south Adriatic margin (outer shelf, slope and basin): a current-dominated system, where, far from direct sediment feeding sources, the oceanography of the basin controls the stacking of stratigraphic units (chapter 6).

The results obtained for each study area have been documented in four research papers (that have either been published, are under peer review, or as in the case of chapter 4 is currently under internal review by Exxon-Mobil researchers for submission to a peer-reviewed international journal) and are presented in chapters 3 to 6. Each chapter consists of a manuscript (edited following the style of the journal) and a brief introduction on the topic of the research and the open questions to which the
manuscript aims to address. In addition, chapter 4 is accompanied by the supplementary material that take into account the age model of PRAD 1-2 borehole done under the guidance of Dr. Alessandra Asioli (IGG-CNR, Padua) and the stratigraphic analysis performed at the Upstream Research Company of the Exxon-Mobil (Houston, TX).

Finally, chapter 7 reports the main conclusion of this dissertation.

The manuscript I (chapter 3),

*River morphodynamic evolution under dam-induced backwater: an example from the Po River (Italy)*

concerns on a portion of the Po River upstream of the Isola Serafini dam (Northern Italy). The trunk of the Po River is viewed as a natural laboratory where investigate and quantify the impact of a backwater zone on river morphodynamics. This analysis helps understanding how the sediment is exported to the ocean and highlights the potential of backwater effect in governing river morphology and sediment partitioning along the channel axis.

The manuscript II (chapter 4),

*Stratal architecture, shelf-edge trajectories and sand bypass on the rapidly accumulated Po River lowstand delta forced by high-frequency base level change*

focuses on the late Quaternary stratigraphic succession of the Mid Adriatic Deep provides the opportunity to investigate with high spatial and temporal resolutions the relation between clinoform progradation and sediment bypass to the basin and refer it to the late-Quaternary global eustatic curve. The high-resolution chronostratigraphy framework of the Po River lowstand wedge provides peculiar insights into shelf-edge architecture and growth phases, which can be applied to models of continental-margin evolution.
The manuscript III (chapter 5),

*Anatomy of a compound delta from the post-glacial transgressive record in the Adriatic Sea*

reports on a supply-dominated system offshore the Gargano Promontory, where the oceanography of the receiving basin regulates sediment dispersion through advection and a markedly skewed thickness distribution of the resulting stratigraphic units. The interplay between sediment supply and sediment advection governed the development of a compound delta with a markedly asymmetric subaqueous clinoform during a short interval of relative sea level still stand during the Younger Dryas cold reversal.

The manuscript IV (chapter 6),

*Pliocene–Quaternary contourite depositional system along the south-western Adriatic margin: changes in sedimentary stacking pattern and associated bottom currents*

investigates a system far from a direct river-born sediment input where the activity of contour-parallel and cascading oceanographic currents govern the formation of the stratigraphic units and their architecture. In particular, at the shelf-edge and upper slope of the south-western Adriatic the interaction between oceanographic currents and the seafloor morphology, which in turn reflects local tectonic deformation, favored the development of contourite systems within the Pliocene-Quaternary succession.
3) Rivers and backwater effect

Sediment flux from rivers to the ocean is the main driver of continental sedimentation with significant implications for land use, construction of hydrocarbon reservoirs, and unraveling Earth history and global climate change from sedimentary strata (e.g., Nittrouer, 1999; Blum and Törnqvist, 2000; Paola et al., 2011). Rivers transfer the sediment to continental margins delivering up to 84% of the total sediment load that reaches the oceans (Milliman and Meade, 1983). River mouths represent fundamental transition zones in the sediment source-to-sink pathway where rivers hand off to marine transport processes. Despite the importance of this transfer, there exists considerable uncertainty about the controls on erosion and deposition of sediment near river mouths (e.g., Fagherazzi et al., 2004). Across the river mouth zone, river flow tends to decelerate downstream due to channel deepening; the segment of river where this response occurs defines the backwater zone (Chow, 1959). At the backwater zone a river behave in fundamentally different ways than further upstream because is affected by the static body of water beyond the shoreline (Fig. 3.1).

![Figure 3.1: cartoon showing a river entering an ocean with three zones of interest: normal flow (x > L), a transitional region (0 < x <L), and the offshore river plume (x < 0) in (a) cross section and (b) plan view. After Lane (1957).](image-url)
Depending on the channel depth at its mouth and on river-bed slope (Paola, 2000), the backwater effect may propagate for tens to hundreds of kilometer upstream, and can influence the gravel-sand and the sand-mud transition of the river (Venditti and Church, 2014), and the river plume hydrodynamics. Together these changes impact the formation of distributary networks and the evolution of delta plains (Jerolmack and Swenson, 2007; Chatanantavet et al., 2012).

The backwater zone can be affected by erosion or deposition depending on river discharge. At low flow the transitional region is a zone of backwater, where the water depth at the shoreline (hs) is greater than the normal flow depth (hn), and the water surface (blue) and bed (black) diverge downstream (i.e., M1 curve [e.g., Chow, 1959]) resulting in deceleration (shown by length of arrows in Fig. 3.1) and deposition. At high flow hn > hs and the water surface (red in Fig. 3.1) is convex (i.e., M2 curve [e.g., Chow, 1959]), resulting in spatial acceleration of flow and erosion. In both cases, the elevation of the water surface at the river mouth is relatively insensitive to discharge due to lateral spreading of the plume (Fig. 3.1b). This concept suggests that backwater dynamics may be important in shaping channel morphology through deposition and erosion (as suggested in Nittrouer et al., 2011; Whipple et al., 2000). Moreover, many backwater zones and estuaries are interpreted to be transient features that result from flooding due to Holocene sea level rise that created accommodation space faster than could be balanced by fluvial sediment infilling (e.g., Anderson and Rodriguez, 2008). This notwithstanding, many rivers have backwater zones that persist despite the formation of Holocene highstand deltas (e.g., Mississippi River, United States).

The next chapter focuses on a portion of the Po River (northern Italy) as a natural laboratory where a backwater zone was created upstream of the construction of the Isola Serafini dam; in the absence of external forcing such as waves or tides avoiding amplification through resonance with waves and tides the area is an ideal study site. By using an integrated approach based on the analysis of geophysical, sedimentological and hydrological data, this study quantifies how river hydrodynamics change through the backwater zone and its impact on river morphology and sediment partitioning along the channel axis. In essence, the backwater and drawdown effects show a
paramount control on the river hydrodynamics and are fundamental in controlling the morphodynamic evolution of rivers, deltas and estuaries. The results obtained may be applied to source-to-sink studies and will help understanding how oscillations in the backwater zone, in response to the allogenic forcing of changing base-level, influence the evolution of channel belts, control sediment export to the ocean and, ultimately, continental margin growth.
3.1) Manuscript I

“River morphodynamic evolution under dam-induced backwater:
an example from the Po River (Italy)”

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Title
River morphodynamic evolution under dam-induced backwater: An example from the Po River (Italy).

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Keywords
Po River, backwater, lateral river migration, gravel-sand transition.

Highlights
The Isola Serafini dam interrupts the Po River continuity at 300 km upstream of its river mouth.
The dam creates a backwater zone that modifies river hydrodynamics for up to 40 km upstream.
Lateral migration rates progressively reduce across the backwater zone.
The size of river-bed sediment decreases across the backwater zone, and coarse-grained channel-bars are progressively drowned and reworked through time.
The dam-induced backwater affected the gravel-sand transition of the river.

Abstract
River dynamics is the product of a combination of autogenic processes and allogenic forcing that may affect the entire fluvial systems, from the source to the downstream end. Among external controls, the backwater effect created by the presence of a standing body of water at the river outlet plays an essential role in the evolution of coastal rivers and delta plains. In natural systems the backwater effect propagates from tens to hundreds of kilometer upstream of the river mouth, and its effect may be amplified through resonance with waves and tides. Here we use a portion of the Po River (Italy) upstream of the Isola Serafini dam, as a natural laboratory to investigate and quantify the impact of a backwater zone on river morphodynamics, where the effect of waves and tides are absent. The analysis of orthophotos and longitudinal cross-sections in pre- and post-dam conditions allows understanding how the Po River adjusts its profile in response to the backwater, quantifying areas of net river bank erosion and deposition in meanders, defining river-bed sediment partitioning. Lateral migration rate is high at the transition from normal to gradually-varied flows, where the river-bed slightly aggrades and bed material is coarser. Moving downstream, across the backwater zone, lateral migration rate progressively reduces; this trend is accompanied by the drowning of channel-bars, by the
reduction of river competence, testified also by deposition of fine grained sediment, and by the increase in bedforms length. These results may be of help in understanding the evolution of natural river systems near the coastlines and in interpreting fluvial-deltaic deposits preserved in stratigraphic sequences.

1. Introduction

Understanding how river systems may evolve in response to external forcing has fundamental implications in predicting the fate of sediment, its final sink and the evolution of continental, shallow-marine and abyssal systems (Pratson et al., 2007). In the last decades an increasing attention has been dedicated to the study of late Quaternary examples, due to the possibility of identifying the signature of tectonic, climatic or eustatic forcing in the fluvial evolution (Blum and Törnqvist, 2000). Moreover, the study of modern systems, with the help of numerical simulations and flume tank experiments, allowed to link the hydrodynamics of the river to the evolution of the deltaic and marine environments, and, ultimately, to the stratigraphic record (Paola, 2000; Van Heijst and Postma, 2001; Edmonds and Slingerland, 2009; Jerolmack, 2009; Kim et al., 2014). An important step forward for more accurate modeling results has been the introduction of a backwater zone upstream of the river mouth, where the flow velocity decelerates until reaching the standing body of water of the ocean (Parker et al., 2008; Hoyal and Sheets, 2009). Numerical simulations and field observations demonstrated that the so-called backwater effect has a paramount control on the evolution of the environments along the sediment routing system, from the continental realm to the open marine. Depending on the channel depth at its mouth and on river-bed slope (Paola, 2000), the backwater effect may propagate for tens to hundreds of kilometer upstream, and may influence the gravel-sand transition of the river (Venditti and Church, 2014), the river plume hydrodynamics (Lamb et al., 2012), and the formation of distributary networks and the evolution of delta plains (Jerolmack and Swenson, 2007; Chatanantavet et al., 2012).

The present study uses a portion of the Po River (Italy) as a natural laboratory where a backwater zone was created by the construction of the Isola Serafini dam (Fig. 1). By using an integrated approach based on the analysis of geophysical, sedimentological and hydrological data, this study quantifies how river hydrodynamics change through the backwater zone, in absence of external forcing such as waves or tides, and its impact on river morphology and sediment partitioning along the channel axis. The results obtained are discussed in the light of understanding modern hydrodynamics of the river with the ultimate aim to provide new insights for the interpretation of ancient deposits and stratigraphic sequences.

2. Study area

The Po River flows roughly West-East from the Western Alps toward the Northern Adriatic Sea, draining an area of ca. 74,500 km² with a 652 km long river (Fig. 1). The watershed can be divided into three sectors on the basis of lithology and maximum elevation: an Alpine sector of crystalline and carbonate rocks (maximum relief ~4500 meters above mean sea level, m amsl), an Apennine sector mostly composed of sedimentary rocks with high clay content (maximum relief ~2000 m amsl) and a central alluvial area including the Po Plain and the Po River delta (Fig. 1).
The annual Po River hydrograph shows two peaks in discharge normally in autumn and spring, generated by rainfall and snowmelt, respectively. The mean annual discharge recorded at the Piacenza gauging station is 959 m$^3$/s (period 1924-2009; Montanari, 2012), while the total annual sediment and freshwater discharges to the Northern Adriatic Sea are $\sim$13x10$^9$ kg and $\sim$40-50 km$^3$, respectively (Syvitski and Kettner, 2007; Cozzi and Giani, 2011).

The Isola Serafini dam, built for hydroelectricity production and completed in the 1962 AD, interrupts the river continuity at ca. 300 km upstream of its river mouth (Figs. 1 and 2). The dam consists of a gate structure with eleven openings of 30 m each that maintain a normal retention level of 41.5 m amsl through a constant flux of water downstream. Depending on the intensity of the floods, the gate can be opened, preventing river overflow and allowing the flux of extra waters. The construction of the dam affected both upstream and downstream river hydrology, morphology, and consequently the flux of sediment, nutrients and dissolved material to the Sea (Davide et al., 2003; Surian and Rinaldi, 2003).

In the 250 km upstream of the dam (watershed of about 45x10$^3$ km$^2$), the Po River course changes from multi-channel braided to wandering to single thread meandering, with a bed slope decrease from 1.4‰ to 0.22‰ and a fining of river bed sediments from coarse gravel to medium/fine sand (Figure 1; Colombo and Filippi, 2010; Lanzoni, 2012). The present study focused on the 70 km of river course upstream of the dam, where channel morphology is characterized by a sequence of meander bends both upstream- and downstream-skewed (Fig. 2). This portion of the river is characterized by a sinuosity index of 1.82, a water surface elevation between 45 m and 41.5 m amsl, a mean and maximum thalweg depth of ca. 6 m and 22 m, respectively (AIPO, 2005). The present-day river bathymetry (Fig. 2) is the result of the interaction between natural processes and anthropic activities, the latter, with a greater impact, in particular related to gravel and sand mining in the years between 1960 and 1970 AD (Lanzoni et al., 2012). The gravel-sand transition has been recognized between the confluence of the Po River with the Tidone and Trebbia tributaries (Fig. 2), where a subtle decrease in slope was created by deformation of the pre-Quaternary substrate (ADBPO, 2005).

3. Material and methods

The morpho-planimetric evolution of the Po River within the study area was quantified by using a combination of orthophotos (years 1954, 1962, 1991 AD) and satellite images (years 2004, 2005, 2014 AD), with the year 1954 AD as a reference for the quantification of river migration rates. Orthophotos were acquired by the IGM (Istituto Cartografico Militare) since the first decades of the 900 Century, and are available online at http://www.igmi.or/voli. Each photo, at 1:33000 scale, was rectified as WGS84 UTM33 N in ArcGIS® by using 150 control points. Satellite images came from Landsat TM (years 2004 and 2005 AD) and Landsat 8 (year 2014 AD), both with 30 m of spatial resolution (data available at http://earthexplorer.usgs.gov). The thalweg depth for the year 1954 AD, used for the numerical simulation of the critical shear stress, came from AIPO (Agenzia Interregionale per il Fiume Po); the data, referenced to mean sea level, are available at http://geoportale.agenziapo.it/cms. The modern thalweg depth came from the multibeam bathymetry of the
river, acquired by AIPO in the years 2004-2005 AD by using a multibeam echosounder Kongsberg 3002 equipped with a DGPS Racal Landstar (Colombo and Filippi, 2010).

Samples of river bed sediment, collected by using a Van Veen grab, and water surface elevation data were acquired during two cruises in July and August 2014, both during low flow conditions (Fig. 3). Positioning and river water surface elevation data were obtained by the using of Trimble Marine SPS461 GPS receiver equipped by an internet-based (via GSM) VRS RTK for real-time corrections that allows to obtain vertical and horizontal resolutions of 10 cm. Water elevation data are normalized for a reference stream height corresponding to a discharge of 1000 m$^3$ s$^{-1}$, measured at the Piacenza gauging station located about 30 km upstream of the dam (data available at http://arpa.emr.it). Elevation data are converted in water discharge by using the rating curve presented in Cesi (2004).

Grain-size analyses were performed with a set of sieves for the >63 mm fraction (from mesh no. 7 to no. 230) and with Micromeritics SediGraph 5120 for finer fractions at ISMAR-CNR sedimentary laboratory. All the samples were treated with 30% diluted H$_2$O$_2$ to remove organic matter and washed with distilled water to dissolve salts. Before SediGraph analyses the samples were dispersed in sodium hexametaphosphate and the flocculation were avoided also with mechanical agitation. Grain-size distributions were interpreted by using the software Gradistat and are presented according to the software graphics (Blott and Pye, 2001).

4. Field observations

4.1. Morpho-planimetric evolution

The morphologic and planimetric evolution of the Po River along the 70 km of river course within the study area have been quantified through orthophotos and satellite images available since 1954 AD, few years before the construction of the Isola Serafini dam. The pattern of the river surface area was obtained through pixel analysis in ArcGIS®, although this approach neglects small planimetric variations associated to the river stage, with larger channel widths during high-flow conditions. The data show that the river course changed continuously through time in the upstream portion, with lateral migrations of few hundreds of meters (Fig. 4). Bend instability progressively reduces downstream, and in the final reaches, from the dam to ca. 30 km upstream, the river almost maintained its position. This tendency of the river is highlighted also by lateral shifts of the river centerline (Fig. 5), positioned with horizontal errors in the order of few meters, depending on the resolution of the images available. Centerline migration rate, low in the most upstream portion of the river, increases between ca. 55 and 30 km, with values up to 45 m/yr in the years spanning the construction of the dam, and starts to reduce at ca. 30 km upstream of the dam, where the river flows close to the city of Piacenza. Migration rates decrease over time along the entire section, until reaching extremely low values in the final reaches of the river (Fig. 5). The condition of river bank stability detectable along the 30 km of river course upstream of the dam persisted during the entire time-interval investigated, between 1954 and 2014 AD.

The planimetric evolution of the river has been accompanied by morphological changes of channel bars, as detected in 4 snapshots between 1954 and 2004/2005 AD (Fig. 4). The most upstream portion of the river, meander loop 1, shows the most energetic conditions, with side bars that start to accumulate between
1954 and 1962 AD (see yellow arrows in Figure 4), most likely after the construction of the dam, and continuously growing since that. The point bar, furthermore, is progressively reworked and reshaped, testifying the dynamism of this trunk of the river (see orange arrows in Figure 4). Moving downstream, meander loop 2, sand transport is progressively reduced, as testified also by the gradual drowning of side bars (between 1962 and 1991 AD, red arrows in Figure 4) and by the slower reshaping of the point bar, with strengthened banks by riparian vegetation (blue arrows in Figure 4). Further downstream, in full backwater conditions, side bars disappear soon after the construction of the dam (between 1954 and 1962 AD in both meander loops 3 and 4, red arrows in Figure 4), whereas sandy point bars are drowned faster close to the dam: between 1962 and 1991 AD in meander loop 3, and between 1951 and 1962 AD in meander loop 4 (red arrows in Figure 4).

5.2. Downstream change in river bed sediment and bedforms

The general trend of downstream reduction in shear stress, emphasized by the morpho-planimetric change of the river, is confirmed by grain size analyses of river-bed material. Bed sediments, collected along the 70 km of river course, with more detailed sampling in meander loops 1, 3 and 4 (Fig. 6), show the presence of coarse-grained material ($D_{50}$ up to 10 cm) only in the upstream portion of the river (along about 40 km): pebble to cobbles river beds develop with a patchy distribution and alternate with sandy beds with mean $D_{50}$ of 600 μm. Pebbles and cobbles accumulate in particular where flow velocity increases, inhibiting the deposition of fine-grained particles (see meander loop 1 in Figure 6), or at the confluence with the main tributaries entering from the South. Here cobbles, characterized by an Apennine provenance, remain in situ without being carried by the Po River (ADBPO, 2005). Cobbles were also found in few samples further downstream, in areas subjected to anthropic interventions for channel bank stabilization. Approaching the meander loop 3, the samples are almost entirely characterized by sand-size particles, with mean $D_{50}$ between 500 μm and 300 μm. Finer deposits, with high clay content, can be found in low-velocity zones, as the inner side of the meander toward the drowned point bar (Fig. 6). Further downstream, within the final reaches of the river next to the dam (meander loop 4, Fig. 6), bed sediment becomes finer and dominated by silt-size fractions, with mean $D_{50}$ between 200 μm and 10 μm. This trend is well highlighted by the grain size distribution of the river-bed sediment sampled in the thalweg along the final 15 km of the river (Fig. 7). A quick comparison between aerial views of the final reaches of the river acquired before and 50 yr after the construction of the dam (1954 AD and 2004-2005, during low flow conditions) highlights clearly how this downstream decrease of particles size is accompanied by the drowning of the system and the disappearance of coarse-grained channel bars. Furthermore, this change is also reflected by an increase in the dimension of river bedforms (Fig. 8). Dune length, in particular, shows a mean value of ca. 5 m in the upstream portion of the river and, starting at circa 30 km from the dam, increases downstream until reaching values up to 25 m.
5. Po River sediment transport capacity

In order to quantify the impact of a backwater zone on river flow velocity, bed shear stress and, ultimately, sediment transport capacity of the river, it is needed to introduce a hydrodynamic model that describes the downstream change of the river surface height approaching a standing body of water.

5.1. 1-D backwater curve

Under the assumption of the conservation of fluid mass and momentum, the depth-averaged gradually-varied flow equation is:

\[
\frac{dh_x}{dx} = \frac{S_0 - S_f}{1 - Fr^2}
\]

with \(h_x\) flow depth, \(x\) downstream distance, \(S_0\) river bed slope, \(S_f\) friction slope and \(Fr\) Froude number (Chow, 1959). Since the flow is only gradually varying, it is possible to calculate the friction slope at any given point (with flow depth \(h_x\)) under the backwater influence by using the Manning’s equation \(U = \frac{1}{n} R_h^{\frac{1}{3}} S_f^{\frac{2}{3}}\). Far upstream from the backwater influence, where the condition of steady and uniform flow is reached, the friction slope is constant and equal to the bed slope (\(S_0 = S_f\)), and the Manning’s equation allows the calculation of the normal flow depth. In rectangular channel (\(H = \text{depth}, L = \text{width}\)), with the approximation for the hydraulic radius \(R_h H\) (as \(L \gg H\)), the normal flow depth is \(H_n = \frac{U^2}{C_z g S_{0 \ell}}\), where the mean flow velocity \(U\) is given by the continuity equation (\(U = Q / HL\)), with \(Q\) is the water discharge, \(L\) is the mean channel length, \(g\) is the gravitational acceleration, \(C_z\) is the Chezy resistant coefficient (Chow, 1959).

For subcritical normal flow regime (\(Fr < 1\)) on mild slope (\(h_1 > H_n > H_c\)), and considering a single discharge with no delta progradation, Eq. (1) can be solved with an iterative procedure starting with the depth \(h_{x=1}\) at the ponded water and integrating upstream (Chow, 1959); the computation allows to have a preliminary estimate of the backwater curve for 1-D open channel flow on a flat sand bed (e.g. Parker, 2004).

The river surface height for pre-dam and post-dam conditions, therefore with a backwater zone, is quantified by using: 1- a constant bed slope of 0.2063 \(10^{-3}\), calculated from the linear fit of thalweg depths derived from cross-sectional channel profiles (Fig. 9); 2- three Chezy resistant coefficient (10 in red, 20 blue, 30 green); 3- a water discharge of 1000 \(m^3 s^{-1}\), derived the mean annual discharge measured at the Piacenza gauging station; 4- a channel width of 250 m, a mean value calculated along the 70 km of the study area; 5- a starting depth of 10 m, as the depth at the normal retention level of the reservoir (41.5 m amsl). Figure 4 compares the normal flow depth for pre-dam conditions (left column) and the M1 river profile for post-dam conditions (center column), calculated for the three Chezy coefficients, to the modern channel depth and river surface height for a discharge of 1000 \(m^3 s^{-1}\) (right column in Figure 9).

5.2. Total boundary shear stress
Once calculated theoretical channel depth in pre and post-dam conditions, and considering the resistance of the alluvial bed only related to the presence of grains, the total boundary shear stress can be quantified using the equation:

\[ \tau_b = \frac{U^2}{Cr_z^2} \]  

(Einstein and Barbarossa, 1952; Wright and Parker, 2004). The channel deepening in consequence of dam-induced backwaters creates an exponential reduction of the shear stress that in turn affects the sediment transport capacity of the river (Fig. 9).

The shear stress calculated for the present configuration of the Po River (right column in Figure 9) is quantified by using the thalweg depth surveyed by AIPO in the years 2004-2005 AD (Colombo and Filippi, 2010). The water surface profile, acquired in this study, shows a sharp decrease in slope from $3 \times 10^{-4}$ to $0.5 \times 10^{-4}$ in the most upstream portion of the study area (between 70 and 45 km), where the gravel-sand transition is recognized (ADBPO, 2005). Further slight changes in slope are in correspondence of the confluence with the main tributaries Tidone, Lambro and Trebbia rivers (Fig. 2).

5.3. Sediment transport capacity

In order to quantify the mobility of specific grain size, it is possible to compare the non-dimensional shear stress (i.e. the Shields parameter $*^c$, estimated from the total boundary shear stress; Bunte et al., 2010) of each grain size to its critical value $c^c$, estimated from Brownlie (1981) as a function of a modification of the Reynolds particle number:

\[ c^c = 0.22 R_p^{0.6} + 0.06 \times 10^{-7} R_p^{0.6} \text{ with } R_p = g s \frac{w_D}{s_D^{0.5}} \frac{D^{0.5}}{D,50} \]  

(2)

with $s = 2650 \text{ kg m}^{-3}$ and $w = 1000 \text{ km m}^{-3}$ particle and water density, respectively, and $= 1.004 \times 10^{-6} \text{ m}^2 \text{ s}^{-1}$ kinematic viscosity at 20 °C.

The downstream change in sediment mobility has been calculated for 12 grain-sizes (Fig. 10), spanning from cobble to silt, and considering a Chezy resistant coefficient of 30, the value that best represents this portion of the Po River, characterized by a clean bed with little to absent vegetation on the walls. The model well estimates the upstream influence of the backwater zone compared to field observations, and well approximates the reduction of the sediment transport capacity of the river for the post-dam conditions. Along the 30 km of river course upstream of the dam, the particle size carried by the river decreases from few millimeters to less than 100 μm; this value is in agreement with the competence of the river calculated for modern conditions, and also with the results from river-bed sediment sampling collected during normal flow conditions.

6. Discussion

The downstream hydrodynamic change of a river entering into the coastal ocean has fundamental implications for the formation and evolution coastal landscapes (Edmonds and Slingerland, 2006; Geleynse et al., 2011). In the last years many efforts have been made to understand how, how much, and how far upstream,
the presence of a standing body of water at a river mouth affects the behavior of the river, its sediment transport capacity, and the morphodynamics of deltas and estuaries (Jerolmack and Swenson, 2007; Edmonds et al., 2009; Chatanantavet et al., 2012; Lamb et al., 2012; Bolla Pittaluga et al., 2015). We used a trunk of the Po River where a backwater zone was induced by the construction of a dam, as a natural laboratory to investigate the morphodynamic evolution of the river under gradually-varied flow conditions, in comparison with the observations made in river systems toward their outlets to the ocean (Nittrouer et al., 2012).

6.1. Morphological change within the backwater zone

When river flow velocity is forced to reduce at the river outlet, the perturbation induced in the river hydraulics migrates upstream for distances that are typically approximated by \( L \approx H/S \) (S: channel bed slope, H: flow depth at the river mouth; Paola, 2000). Throughout the backwater zone, the morphology and kinematics of the Po River changes downstream, reflecting the non-linear reduction in the sediment transport capacity of the river. At the transition from normal to gradually-varied flow conditions under the backwater influence (between 45 km and 30 km, approximately), the lateral migration rate is high and accompanied by a slight tendency of the river to aggrade (Figs. 2 and 5), in marked contrast with the scenario further upstream, where river incision may be the consequence of reduced sediment supply from tributaries and river-bed mining. This tendency of the river, explored also by the Exner model results in Nittrouer et al. (2012), reflects the spatial divergences in river-bed sediment transport associated to non-uniform water flow, and has fundamental implications for the onset of distributary network on deltas (Edmonds et al., 2009; Chatanantavet et al., 2012).

Along the final reaches of the Po River, downstream of 30 km and, in particular, close to the Isola Serafini dam, lateral migration rate collapses to extremely low values, sandy channel bars disappear and the river progressively erodes its bed along the thalweg. This tendency of the river has been observed also in the Mississippi River, where lateral migration rate decrease from more than 120 m/yr to less than 20 m/yr along the last 600 km upstream its outlet (Hudson and Kesel, 200). To fully understand the morphodynamic of this trunk of the Po River it is necessary to evaluate the hydrodynamic conditions during both normal flow and high-discharge events. In the first case, a reduction in flow velocity, typical of M1 river profiles, corresponds to a drop of the bed shear stress below critical thresholds for both suspended-load and bed-load transport and to the accumulation of fine-grained material, up to clay size, in the channel axis and along channel bars. During floods and high-discharge events, conversely, the channel-bed kinematics change drastically: when the normal-flow depth of the river becomes larger than the water depth at its downstream end, and this is because the normal retention level of the dam is maintained at a constant elevation by opening the gates, the river flow velocity accelerates due to the drawdown of the fluvial water close to the dam (as described by an M2 profile), the shear stress increases and the river becomes erosional. This phenomenon, first observed by Lane (1957) and quantified by Lamb et al. (2012), explains the reduction in lateral river migration, the deep incision of the channel thalweg, that may promote the formation of armored beds, and the partial removal of the sediment accumulated in channel bars and in the final reaches of the river. Similar processes have been observed in the Teshio River (Ikeda, 1989) and in the Mississippi River (Nittrouer et a., 2012); in the Lower Mississippi River,
in particular, a 100-fold increase in sand transport has been observed during high-discharge events, when the
drawdown of fluvial water forced an up to 13-fold increase in shear stress respect to backwater conditions
(Nittrouer et al., 2011). The decrease in lateral migration, favoured by the stabilization of river banks through
riparian vegetation (Wickert et al., 2013), coupled with the deepening of the channel toward its outlet, promotes
the narrowing of the channel belt. This concept may be useful when interpreting ancient fluvial systems, as
low values of the width-thickness ratio of the channel belt associated to limited lateral migration rate (up to
few channel widths) may be diagnostic of distributary channels (Blum et al., 2013).

6.2. Sedimentological change within the backwater zone

The hydrodynamic behavior of the Po River throughout the backwater/drawdown zone can be
quantified not only by considering spatial informations such as lateral river migration, channel deepening and
widening, but also by considering the morphological evolution of channel bars and the grain-size partitioning
along the channel axis. Repeated aerial surveys of the Po River during the last 60 years, since dam-induced
backwater has been created, show that point bars and side bars, continuously evolved through time between
70 and 45 km (Fig. 4), as the river-bed shear stress is high enough to transport coarse sands also during
moderated flow conditions (1000 m³ s⁻¹, Figure 10). Further downstream, and in particular along the 30 km of
river upstream of the dam, the establishment of a backwater zone forced channel deepening, while repeated
oscillations backwater/drawdown associated to the transit of high discharge events progressively eroded the
channel bars and further increased thalweg depth (Figs. 2 and 4). The morphological changes associated to the
kinematics and transport capacity of the river is reflected in channel-bed grain size data: coarse sediment
characterizes the upstream portion of the river and accumulates in the trunk of the river where the transition
from normal-flow to gradually-varied flow is observed (between 45 km and 30 km, approximately), and that,
interestingly, corresponds to the gravel-sand transition of the river. Further downstream pebbles and coarse
sand are gradually replaced by fine sand, silt and, close to dam or where the flow velocity is extremely low,
by clay. River-bed samples from the drowned point bars in meander loops 3 and 4 (Fig. 6), in detail, show a
coarsening trend toward deeper water, with clay-rich sediment accumulating at the inner point bar (Fig. 6).
This situation is in marked contrast with the morphology of both meanders observed in 1954 AD, before the
dam, where sandy point bars are subaerially exposed (Fig. 7). On the basis of the lithology observed in meander
loops 3 and 4, it is possible to speculate that a fining upward trend, with more frequent mud-drapes, should
characterize the vertical sedimentary succession of a point bar where the river enters the backwater zone. As
the backwater zone penetrates further upstream respect to the tidal influence and to the intrusion of brackish
waters (Blum et al., 2013), this finding may be useful when interpreting ancient river deposits at outcrop or
sediment core scales, where the presence of fine-grained beds are normally associated to tidal influence
(Shanley et al., 1992; Labrecque et al., 2011).

The change in river hydraulics and river-bed material observed throughout the backwater zone is
reflected also in the evolution of river bedforms, showing a downstream increase in length (Fig. 8). This trend
may be connected to a combination of the general fining of bed sediments and the high flow velocity during
high-discharge events, as both factors contribute positively to increase the size of the dunes (Southard and Boguchwal, 1990). Therefore, it is possible that only large-scale dunes formed during high-discharge events and remain as relict features (i.e. not in equilibrium) during low flow conditions, as observed in the Yangtze River (Chen et al., 2012).

6.3. Implication for the gravel-sand transition

It is widely recognized that the size of bed sediments in alluvial rivers decreases exponentially downstream (Church and Kellerhals, 1978), and among such variation the transition from gravel to sand is a fundamental boundary in the fluvial geomorphology, as it marks the key change from gravel-bed to sand-bed rivers. Several mechanisms, both autogenic and allogenic, have been proposed to explain the emergence of a gravel-sand transition (Schumm and Stevens, 1973; Hoey and Ferguson, 1994; Ferguson et al., 1996; Ferguson, 2003), and among external controls the backwater influence on river capacity plays a fundamental role (Sambrook Smith and Ferguson, 1995; Venditti and Church, 2014). The gravel-sand transition in the Po River has been recognized between the confluences with the Tidone and Trebbia tributaries, where river hydrodynamics start feeling the effect of backwater (Fig. 2), and has been associated to a slight decrease in river bed slope due the deformation of the pre-Quaternary substrate (ADBPO, 2005). Combining the morphological evidences of the Po River derived from the orthophoto acquired in 1954 AD (Fig. 7), showing the presence of large point bars, side bars and both vegetated and un-vegetated mid-channel bars, with the sedimentological informations available within the study area and further downstream of the Isola Serafini dam, where armored gravel beds have been recognized below the modern sedimentary cover (ADBPO, 2005), it is possible to argue that, before the construction of the dam, the gravel-sand transition was located kilometers downstream respect to the modern position. Therefore, the modern location of the gravel-sand transition is not as it would be for a pristine system, but it is consequence of the upstream effect of dam-induced backwater. This result adds value to the importance of the backwater effect in controlling not only the evolution of the downstream end of the river, with its impact on the genesis of distributary network on deltas and on the morphodynamics of estuaries (Jerolmack and Swenson, 2007; Edmonds and Slingerland, 2009; Chatanantavet et al., 2012; Lamb et al., 2012; Bolla Pittaluga et al., 2015), but also the kinematics and morphology of the river for kilometers upstream.

7. Conclusions

The hydrodynamic behavior of a river changes radically from its mountain reaches to downstream toward the ocean. The presence of a standing body of water at the river outlet exerts a fundamental control on the river morphodynamics through the formation of a backwater zone that may propagate for hundreds of kilometers upstream. Quantifying the morphologic evolution of a trunk of the Po River when entering in a dam-induced backwater zone allows understanding the effect of the transition from normal to gradually-varied flow regimes on the river dynamics, without introducing assumptions on the effect of waves and tides. From upstream to downstream the lateral migration rate of the river reduces, accompanied by a general fining of bed
sediments and increase in dune length. When the river enters the backwater zone, a strong decrease in water surface slope and associated bed shear stress allows the deposition of coarse-grained material, promoting the gravel-sand transition of the river and the aggradation of the river bed. Further downstream, where the water surface oscillates from M1 to M2 profiles during low-flow and high-discharge events, respectively, bed shear stress increases promoting the erosion of channel bars and the deepening of the channel thalweg. In essence, the backwater and drawdown effects show a paramount control on the river hydrodynamics and are fundamental in driving the morphodynamic evolution of rivers, deltas and estuaries. This concept, applied to source to sink studies of continental margin evolution, will help understanding how oscillations in the backwater zone in response to the allogenic forcing of changing base-level govern the evolution of channel belts, control sediment export to the ocean and, ultimately, continental margin growth.

Acknowledgments

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Figure 1. Po River course and its catchment basin, with reported the main tributaries entering the river in the study area (black box). River bed elevation and mean slopes, calculated for the 250 km of river course upstream of the dam (between point A and B; modified from Colombo and Filippi, 2010). Point B marks the location of the Isola Serafini dam.
Figure 2. Top: Aerial view of the study area derived from a combination of two images (date 7/21/2004 and 5/8/2005, left and right sides of the white dashed line, respectively) acquired during low flow conditions, see Figure 3 for details. Red lines represent the location of river cross sections acquired in 1954 AD (source AIPO: http://geoportale.agenziapo.it/cms). Bottom: thalweg elevation from multibeam bathymetry (black line; source AIPO, 2005) and from 1954 AD river cross sections (squares and numbers reported in red). Blue arrows are the four main tributaries (Tidone, Lambro, Trebbia and Nure rivers). The blue star marks the location of the gauging station near the city of Piacenza.
Figure 3. Top: Water surface elevation (blue) and water discharge (red) data from the Piacenza gauging station (source: http://arpa.emr.it) during June-September 2014; cruise time is reported below. Bottom: water surface elevation of the Po River during the acquisition of the aerial views presented in Figure 2; both images are acquired during low flow conditions (black squares).
Figure 4. Top: Extent of the Po River surface area obtained from five aerial surveys since 1954 AD (before the construction of the Isola Serafini dam) highlighting the downstream change in lateral migration of the river. Bottom: The detailed view (age reported in years AD) of four selected meander loops (1 to 4 moving downstream) shows the morphological evolution of river banks and channel bars. In detail: orange and yellow arrows in meander loop 1 indicate channel bar and point bar accretion through time, respectively; blue arrows in meander loop 2 mark river banks with increasing vegetation cover; red arrows in meander loop 2, 3 and 4, highlight channel bars that are progressively eroded until disappear.
Figure 5. Lateral migration rate of the Po River centerline, calculated in four time intervals, plotted versus distance upstream. The data show a marked drop of the lateral migration of the Po River along the 30 kilometers upstream of the Isola Serafini dam, in the interval 1954-1962 AD. During the last sixty years lateral migration reduce both in space (downstream) and in time (toward nowadays).
Figure 6. Detailed view of meander loops 1, 3 and 4 (see location in Figure 4) showing the grain-size variability of river bed sediment. Blue (sediment < 63µm), yellow (sediment between 63-2000 µm) and red (>2000 µm), with color proportions reflecting the amount of a specific grain size. Note the overall decreasing in grain size toward the inner side of the meander loop 3 and approaching the Isola Serafini dam.
Figure 7. Morphology of the Po River along the final reaches located upstream of the Isola Serafini dam in 1954 AD and 2005 AD. Note the presence of extensive coarse-grained deposits along the river course before the establishment of backwater conditions in marked contrast with the modern river channel physiography lacking of channel bars. River-bed samples acquired in this study (colored dots reported on the 2005 AD satellite image) highlight the general downstream fining of sediment particles, as summarized by the grain size distributions (each color refers to a specific sediment sample).
Figure 8. Variability of river bed sediments along the 70 kilometers of the Po River upstream of the Isola Serafini dam (blue dot, dimensions scale with D<sub>50</sub> grain sizes). Vertical blue bars represent the range of grain sizes, while horizontal bars their spatial distribution along the river. Red diamonds represent average dune length derived from 300 m long 2D profiles extracted from multibeam bathymetry. Vertical red bars represent the maximum and minimum dune length in each transect.
Figure 9. Theoretical water surface elevations (w.s.e.) for normal flow (pre-dam) and gradually-varied flow (post-dam) conditions, calculated by introducing three Chezy coefficients (10 in red, 20 in blue, 30 in green). Water surface slope is equal to the bed slope (obtained through linear fit of the thalweg depths derived from river cross-sectional profiles, see Figure 2) in the normal-flow regime, while is derived from the M1 river profiles in the post-dam conditions. Total boundary shear stress ($\tau_b$) is by using a channel depth derived from river bed elevation (r.b.e.) and theoretical water surface profiles, and shows the gradual downstream decrease in response to the establishment of dam-induced backwater. The right column represents the modern river physiography, with thalweg depth in black and water surface elevation in blue; the water surface slope in red; and, below, the total boundary shear calculated by using a water discharge of 1000 m$^3$ s$^{-1}$. The drop in shear stress, between kilometers 45 and 30, coincides with the upstream limit of the backwater zone.
Figure 10. The sediment transport capacity of the river has been quantified through change in frictional mobility estimated by comparing the non-dimensional shear stress of specific grain sizes to its critical value. The post-dam scenario, with a marked downstream reduction of the ratio $t^*/t_c^*$ compared to pre-dam conditions, is in agreement with the modern river transport capacity. River bed profile in pre- and post-dam scenarios are obtained from thalweg depths in river cross-sectional profiles; the modern river profile is derived from the 2005 AD multibeam bathymetric survey (Colombo and Filippi, 2010) while water surface elevation are acquired in this study and scaled for water discharge of 1000 m$^3$ s$^{-1}$. 
4) Clinoforms and trajectory analysis

Clinoform is the term originally introduced by Rich (1951) to describe the shape of a depositional surface at the scale of an entire continental margin (Fig. 4.1). Clinoforms are widely recognized as one of the fundamental building blocks of the stratigraphic record (Pirmez et al., 1998) and are characterized by three geometric elements: the topset (undaform) represents the most low-angle and shallow-water sector; the foreset (clinoform) diagnostic of the steepest sector; the bottomset (fondoform) the almost flat and distal basinward sector (Fig. 4.1). The morphological break in slope at the topset-foreset transition is called rollover-point and in margin-scale clinoforms it coincides with the shelf edge (Fig. 4.1).

![Figure 4.1: Diagram showing the original definition of clinoform from Rich (1951). Note the position of subaerial and subaqueous rollover points.](image)

Since the advent of seismic stratigraphy, clinoforms have been observed over several spatial and temporal scales ranging from shoreline accretion to continental margin progradation (Vail et al., 1991), and encompassing intervals from hundreds of thousands to millions of years (Helland-Hansen and Hampson, 2009). Further advances in understanding of the migration of clinoforms through time derive from the application of the rollover-point trajectory analysis (Fig. 4.2; Helland-Hansen and Martinsen, 1996). The trajectory analysis is the study of lateral and vertical migration of geomorphological features and associated sedimentary environments, with emphasis on the paths and direction of migration of the shoreline and shelf edge (Henriksen et al., 2009). This information can
be used to gain enhanced understanding of changing paleoenvironmental conditions and lithological distributions through time.

Figure 4.2: schematic diagram showing the different prograding shoreface and shelf clinoform. Note that the successive positions of the shoreline allows for the identification of a shoreline trajectory, which, in this example, shows normal regressive trend. The shelf edge trajectory is determined by the successive positions of the migrating shelf-slope break and may exhibit a ascending, flat or descending trend (modified after Steel and Olsen, 2002; Patruno et al., 2015).

More precisely, clinoform trajectory analysis has been applied to constrain the migration pattern of shorelines and associated coastal depositional systems (Helland-Hansen and Martinsen, 1996), or to describe the cross-sectional pathway of the shelf-edge during the accretion of shelf-slope-basin
clinoforms (Mellere et al., 2002; Johannessen and Steel, 2005; Mahon et al., 2015). Trajectory analysis has been mostly applied to two-dimensional (2D) studies of widely-exposed dip-oriented successions, both in outcrop (Mellere et al., 2002, among others) and on seismic profiles (Patruno et al., 2014, among others), to progradational successions encountered in the geological record and ascribed to base level cycles of 100-500 kyr duration (Hampson, 2010) or even lower frequency (Midtkandal and Nystuen, 2009). The integrated approach of trajectory analysis and traditional sequence stratigraphy allows a better understanding of how changes in base level, subsidence and sediment supply affected the depositional architecture of continental margins (Fig. 4.2). While the shoreline trajectory is determined by the interplay among sediment supply, 4th and 5th order sea level changes, and basin physiography (Helland-Hansen and Gjelberg 1994; Helland-Hansen and Martinsen 1996), the shelf-edge trajectory seems more influenced by lower order changes of relative sea level and sediment flux (Mellere et al., 2002; Bullimore et al., 2005; Helland-Hansen and Hampson, 2009). Earlier publications suggest that descending/flat shelf-edge trajectories are associated with more fluvial-dominated depositional systems and that ascending shelf-edge trajectories are associated with wave-dominated processes; in addition, flat/negative shelf-edge trajectories have coeval deepwater slope and basin floor sandy deposits, whereas positive trajectories tend to associate with muddy slopes and basin floors coupled with the sand storage on the shelf and coastal plain (Steel el al. 2000; Plink-Björklund et al. 2001; Mellere et al. 2002; Plink-Björklund and Steel 2002; Bullimore et al., 2005).

The next manuscript is focused on the integrated approach of traditional stratigraphy with shelf-edge trajectory analysis of the Po River Lowstand delta that developed during the Last Glacial Maximum in the central Adriatic Sea. The additional material that accompanies the manuscript document the age model built for the interested stratigraphic interval through the analysis of the PRAD 1-2 borehole (paragraph 4.1.2) and the stratigraphic analysis of one of the clinothems that recorded the formation of deep-sea sandy lobes (paragraph 4.1.3). The stratigraphic interpretation takes into account surface, isochronal and seismic facies maps of each elementary clinothems that
were analyzed through Petrel 3D platform at the Upstream Research Center (ExxonMobil, Houston, Texas).

This study offers the opportunity to resolve the internal geometry of clinothems that compose a lowstand sediment wedge and verify the presence of a systematic depositional motif. The high spatial and temporal framework gives helps establishing predictive relationships between shelf edge trajectories and the occurrence of sand-rich slope deposits.
4.1) Manuscript II

“Architectural motif, shelf-edge trajectories and sand bypass in short lived low-stand clinothems: the late-Pleistocene Po delta”

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Architectural motif, shelf-edge trajectories and sand bypass in short lived low-stand clinothems: the late-Pleistocene Po delta

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Abstract

The Late Pleistocene Po River lowstand delta rapidly filled the Adriatic foredeep through a 350 m thick set of longitudinally prograding clinothems, causing a margin encroachment of 60 km during an extremely short time window, between 31.8 and 14.4 kyr BP. The resulting very expanded stratigraphic succession allows investigation of the spatial relations and internal architecture of each composing clinothem, in a high-resolution chronostratigraphic framework constrained by a continuous-recovery borehole (PRAD1-2) in the distal bottomset area. Based on topset geometry, shelf-edge trajectories and seismic facies, three types of clinothems can be recognized, each associated with different amounts and mechanisms of sand transport offshore. In Type 1, characterized by descending shelf-edge trajectories and eroded topset, sand is transported to slope channels and bottomset lobes; in Type 2, associated with ascending trajectories and topset aggradation, sand is transported beyond the shelf-edge by thin-skinned mass failures; in Type 3, characterized by ascending trajectories, maximum topset aggradation and major landward shifts of the coastal onlap, sand is almost entirely trapped on the shelf, and can reach the basin through hyperpycnal flows of decreasing competence through time, as recorded by cm thick sharp-based sandy layers on the slope. While the alternation of Type 1 and 2 clinothems, formed during overall sea level falls of about 25-30 meters, is associated with high-frequency accommodation cycles (mostly driven by the compaction of rapidly-accumulated sediment), Type 3 heralds the abandonment of the system close to the LST to TST transition (onset of Terminantion I), when sand transport to deeper water is progressively hampered by enhanced alluvial plain aggradation.

Our findings and chronological constraints document that subsiding foredeep settings at lowstand may record the typical stack of parasequences encountered in other subsiding realms, like the
Cretaceous successions of the Book Cliffs, but with durations one to two orders of magnitude shorter. Moreover, in over-supplied basins like the Po, repeated high-frequency accommodation cycles of only few thousands years duration may lead to the stacking of disconnected sand bodies in slope and bottomset settings, resulting in compartmentalized reservoirs even within one single Milankovitch-scale lowstand wedge.

**Introduction**

Clinothems are the major building block of continental margins and their variable morphological profile reflects the interplay between sediment supply and available accommodation space (Pirmez et al., 1998). Being the shape of a clinotheme somehow independent from its physical scales, attempts have been made to define a hierarchy of stratal patterns and key surfaces to interpret the conditions for clinotheme growth virtually on any time scale (Neal and Abreu, 2009). A major advance in understanding the evolution of clinothems and continental margin growth came from the integration of sequence stratigraphy with the trajectory analysis of morphological breaks-in-slope, often referred to as rollover points in the literature (i.e. the topset-foreset transition; Steel and Olsen, 2002). This integrated approach led to a better quantification of the temporal range of creation/destruction of accommodation space and its bearing on the prediction of lithology and facies partitioning at a sequence scale (Henriksen et al., 2009 and references there in).

The Po River lowstand delta (ca. 200,000 km$^2$ drainage area debouching on a ca. 20,000 km$^2$ basin; Fig. 1), accumulated in a over-supplied epicontinental sea during the Last Glacial Maximum (LGM), offers the unique opportunity to disentangle how allogenic and autogenic processes drive clinotheme growth, sediment export toward the slope/basin and sand bodies compartmentalization within a single progradational event and with a very high chronological resolution.
Results

The interval between the former interglacial (Eemian, 123-119 kyr BP) and the onset of the LGM at ca. 23 kyr BP (LGM Chronozone Level 1, 19,000-23,000 cal yr B.P. according to Mix et al., 2001) led to a substantial shrinkage of the Adriatic basin and the concurrent broadening of the Po River drainage area, with a seaward shift of the shoreline of ca. 200 km (Maselli et al., 2011; Amorosi et al., 2015 and Fig. 1). During the LGM, the Po River discharged directly into the Mid Adriatic Deep (MAD), a structurally-confined slope basin that remained connected to the Mediterranean Sea, thereby keeping a link to global sea level.

The chronology of the progradation into the MAD was assessed in the borehole PRAD1-2, recovered in its distal toe region (Ridente et al., 2009), by a combination of new and already available $^{14}$C AMS dates (Piva et al., 2008), tephrochronology (Bourne et al., 2010), bio- and event-stratigraphy (Piva et al., 2008; see Supplementary material for details) (Fig. 2): the delta started its growth at ca. 31.8 cal kyr BP and was drowned at ca. 14.4 cal kyr BP, when a major flooding event brought the basin into fully marine waters (Asioli et al., 2001). In this very short time-window, encompassing the MIS 2 sea level fall (about 25-30 meters, Siddal et al., 2003) to the early stages of the post-glacial sea level rise, the Po River outbuilt its coastal plain 60 km basin-ward, forming a lowstand delta that reached a thickness of 350 m, an estimated total volume of 650-700 km$^3$ (Amorosi et al., 2015; Fig. 3), where 45 m of topset aggradation resulted from the stacking of of coastal/delta plain and paralic deposits.

In detail, clinothems are characterized by repeated changes in rollover-point trajectory accompanied by concurrent changes in the distance between the coastal onlap and the correlative rollover point (Fig. 3). Based on seismic facies, topset geometry, rollover-point trajectory, and occurrence vs. absence of mounded deposits in the bottomset (taken as a proxy of sand bypass across the shelf and slope), three types of clinothems are recognized (Fig. 3 and 4).

Type 1: topset characterized by truncated toplap reflectors that pass vertically to seaward dipping shingled reflectors reminiscent of shelf-phase delta (Suter and Berryhill, 1985). These deposits are connected to discontinuous high amplitude and mounded reflector packages in the bottomset
interpreted as sand lobes; occasionally, these lobes show small channels with modest levee wedges. These clinothems show minimal aggradation in the topset and a descending trajectory of the rollover point (Fig. 3 and 4); in such context it may prove difficult to estimate the amount of sediment removed from the topset region and, ultimately, quantify the relative sea level fall intervened.

Type 2: topset characterized by continuous high amplitude reflectors that pass toward the bottomset into discontinuous and low amplitude reflectors, interpreted as mass-transport deposits. Type 2 clinothems show an ascending rollover point trajectory and maximum aggradation in the order of 10-15 m, calculated through the vertical component in the rollover point trajectory (Fig. 3 and 4). Type 3: topset characterized by plane-parallel landward-onlapping reflectors with high lateral continuity accompanied by weakly reflective foreset with continuous reflectors at the toe region. Type 3 clinothems show aggradation at the topset with a maximum component at the rollover point (10-15 m), an ascending rollover point trajectory and a maximum distance between the landward pinch out and its time equivalent rollover point (Fig. 4).

Type 1 clinothems pass continuously, with no detectable break in sedimentation, into Type 2 clinothems, forming a typical progradational-degradational to progradational-aggradational sequence (APD-PA, Neal and Abreu, 2009). The end of a Type 2 progradation is marked by a flooding surface that has a clear physical expression in seismic profiles and can be traced to borehole PRAD1.2. The seismic facies configuration and the sediment cores (Fig. 2) sampled from Type 3 documents the presence of sharp-based cm-thick sandy layers in the foreset, even if the system is aggrading on a large shelf portion, while sand lobes in the bottomset are absent.

Overall, the height of each clinothems (vertical distance between rollover point and the inflection point at the toe of the foreset; Pirmez et al., 2008) decreases from the oldest (260 m) to the youngest (80 m) during the development of the Po River lowstand wedge; this trend reflects the progressive fill of the basin through continuous bottomset aggradation, in turn favored by the structural confinement of the slope basin (Fig. 3).
In less than 15 kyr, the Po River lowstand delta recorded three APD-PA sequences (three couples of Type 1+Type 2 clinothems), each separated by a minor flooding surface; this succession is topped by a set of Type 3 clinothems (Fig. 4). The first sequence formed in about 10 kyr (between 31.8 and 21.1 kyr BP) and is characterized by the lowest values in the ratio of Ca/Ti obtained from borehole PRAD1-2 (Fig. 2), indicating a reduce influence of the catchment on the sediment supply. The bulk of the progradation (sequences 2 and 3) occurred between 21.1 and 18 kyr BP and is accompanied by higher Ca/Ti values implying an increasing supply of clastic sediment from the Alps and Apennines during the LGM chronozone (an increased biogenic production is excluded by the very low content of carbonate biosomes); furthermore, the presence of the benthic foraminifera *Sigmoilina sellii* throughout the entire section confirms that this phase of the progradation occurred during an overall cold period, while the increasing content of organic matter in PRAD 1.2 suggests hypoxic to anoxic conditions at the sea floor (Fig. 2). The Type 3 clinotheme started accumulating at ca. 18 kyr BP (at ca. 6.5 m in PRAD1-2), during the early phases of the post-glacial sea level rise; the drowning and abandoned of the system at ca. 14.4 kyr (at ca. 2.5 m in PRAD1-2) is accompanied by an increasing rate in the number of benthic specimens and a lighter $\delta^{18}$O values (*Bulimina marginata*), and the last occurrence of *S. sellii* within the uppermost strata (Figs. 2 and 4).

**Discussion and Conclusion**

The Po River lowstand delta is composed by a combination of forestepping clinothems that mimic those commonly ascribed to base level cycles of 100-500 kyr of duration or even lower frequency (Helland-Hansen and Martinsen, 1996; Johannessen and Steel, 2005; Helland Hansen and Hampson, 2009; Midtkandal and Nystuen, 2009; Hampson, 2010). In addition, the descending trajectory of Type 1 clinothems composing the Po River lowstand delta are accompanied by sand leakage through the coastal system and bypass to the upper slope area and are diagnostic of the presence of significant storage of sand in the bottomset, similarly to what is suggested for clinothems that develop on longer time scales (Steel et al., 2000; Gong et al., 2015). Our data therefore show that the geometry and the
thickness of prograding clinotheme packages are not sufficient to discriminate their hierarchy and the
duration of the cycle during which deposition has occurred. As observed originally by Boyd et al.
(1989) for the entire Mississippi River lowstand delta, also in the case of the Po River lowstand delta
the physical scale of the lithosomes recorded in the stratigraphic succession has no implication on the
time elapsed during their deposition.

Although, the formation of the Po River lowstand delta was driven by the allogenic control of eustasy
that was also responsible for part of the topset aggradation, in detail its stratal architecture and facies
distribution were dictated by a combination of changes in sediment supply, mostly driven by climatic
oscillations (as suggested elsewhere by Fuller et al., 1998), and high-frequency accommodation
cycles, related to the compaction of rapidly accumulated sediment. Three APD-PA sequences formed
during an overall sea level fall, as witnessed by the descending trajectory of Type 1 clinothems, but
among each sequence the compaction of Type 1 clinothems results in increased accommodation space
on the shelf that allows the formation of the PA reflector packages within the overlying Type 2
clinothems, indicating that sediment supply was able to keep pace with 5-10 m base-level rises.
Recent estimates of the compaction rates from the rapidly accumulated (30 m in 400 yr) modern Po
River delta (Teatini et al., 2011) are in agreement with the results obtained from the PRAD1-2
borehole, where sediment compaction is calculated in about 15% of the total vertical section (Maselli
et al., 2010). These figures appear consistent with the accommodation space needed for the topset
aggradation of each PA sequences. The formation of Type 2 clinothems is also characterized by the
presence of mass-transport deposit in the bottomset, probably generated by the high accumulation
rates at the shelf edge. At the end of the cycle, rapid basin-ward and downward shifts of the shoreline
generate a bypass surface on the shelf (where accommodation space is at a minimum) and to the
formation of Type 1 clinothems in the basin, where the descending rollover point trajectories reflect
the general sea level fall (Fig. 4). During the formation of Type 1 clinothems, sea level fall forced the
shoreline close to the shelf edge, allowing a more efficient sand transport toward the basin with the
formation of channels and lobes. Changes in sediment accumulation rate obtained from the borehole
PRAD1-2 may reflect climate variability, as indicated by phases of Apennine and Alpine glaciers advances and retreats (Giraudi, 2011; Florineth and Schluchter, 1998), and variations in the sediment storage and bypass toward the basin related to the fluvial response time and autogenic processes (Muto and Steel, 2002; Nijhuis, et al., 2015). The finding from the Po River lowstand delta highlights that, although coarse-sediment delivery to the deep sea is expected principally during periods of sea level fall and lowstand (Covault and Graham, 2010), sand bypass toward the basin may not be continuous even on very short times: high-frequency base level oscillations may force the development of sandy deposits separated by flooding surfaces and relatively thick muddy clinothems (Fig. 4). In this scenario, stacked sandy deposits separated by mud drapes constitute discrete-individual sedimentary bodies and may represent an interesting case of compartmentalization within a single high-frequency lowstand deposit (Ainsworth, 2010).

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Fig. 1: A) Study area with ISMAR high-resolution seismic data (gray lines) and the location of multichannel seismic lines (note the location of seismic line LSD22 in red) and PRAD1.2 borehole (large green dot). B) Po River drainage area during LGM (modified after Maselli et al., 2011) and Alpine and Apennines glaciers extent (modified after Florineth and Schlüchter, 1998; Giraudi, 2011). White lines represent the ancestral Po River and its tributaries; Green and yellow polygons depict Po River coastal plain and delta front, respectively.
Fig. 2: Core images illustrate respectively, 1: the sharp-based sand beds in the proximal foreset and 2: the bioturbated organic-rich etherolitic sediment of the bottomset of Type 3 Clinotheme; in Type 3 the “healing phase” uniform mud records the last 14.4 kyrs (Asioli et al., 2001). Borehole PRAD1-2 in the distal correlative (right) allows dating individual flooding surfaces that partition the expanded lowstand wedge of the LGM (left); $^{14}$C constrained stable isotope curves on planktonic and benthic foraminifera allow correlation to global sea-level curves. Even in the distal location of borehole PRAD1-2, peaks of supply occur consistently during the formation of Type 1 Clinothems, characterized by descending trajectory and sand bypass to the slope basin.
Fig. 3: Multichannel seismic profile LSD22 and line drawing (bottom left). Subtle changes in the topset geometry and shelf-edge trajectories reflect changes in accommodation accompanied by changes in the foreset and bottomset stratal geometry, primarily related to the amount of sand bypass off shelf. Above the sequence boundary (at ca. 31 kyr), individual clinothems show a decreasing height (from 260 m to 80 m) and seaward extent (18.5 km to 3.5 km) as the bottomset area aggrades substantially filling the slope basin and rising significantly the foundation surface for the clinothems. The youngest clinotheme, above the green reflector and records a major landward shift of the shoreline accompanied by high supply to maintain shelf-edge progradation and occurs just before 15.7 kyr BP.
Type 1 and Type 2 alternate repeatedly. Type 1 records the destruction of accommodation (expressed by the erosional topset surface), associated with sandy lobes in the bottomset, and is capped by flooding surfaces that record landward shifts of the coastal onlap in the order of tens of kilometers. Type 2 records constructional accommodation phases (as suggested by ascending rollover point trajectories) coupled with the presence of mass transport deposits in the basin. Type 3 develops at the end of the lowstand and heralds the drawn of the system with maximum aggradation rates on the shelf and the greatest landward migration of coastal onlap; even in these conditions cm-scale sand layers reach the slope as hyperpycnal flows (see Fig. 3).
4.2) Additional Material

4.2.1) Chronology of the PRAD 1-2 borehole

The chronology of the investigated interval of borehole PRAD1-2 relies on the integration of both new and already available in literature (Table 4.1; Piva et al., 2008) $^{14}$C AMS dates, accompanied by tephrochronology on macro and cryptotephra (Bourne et al., 2010) as well as on bio- and event-stratigraphic control points (Piva et al., 2008). The six radiocarbon dates available from literature were performed on benthic monospecific samples, on either *Elphidium crispum* or *Hyalinea balthica*, at the Poznan Radiocarbon Laboratory, Poland, while the five new ones were performed on benthic monospecific samples, on either *Elphidium crispum*, *Glandulina laevigata* at the NOSAMS National Ocean Sciences Accelerator Mass Spectrometry Facility, Department of Geology and Geophysics, Woods Hole Oceanographic Institution, USA and at Poznan Laboratory. The specimens were picked up from the size fraction >0.250 mm.

All the ages were calibrated for the present study using the online Calib 7.1.0 Radiocarbon Calibration Program (Stuiver and Reimer, 1993) and the calibration data set Marine13.14c by Reimer et al. (2013). $\Delta R$ (reservoir) of 135.8 years with a standard deviation of 40.8 years, was obtained by two sites on the western side of the Adriatic: one from the Northern Adriatic (487 years, Rimini) and another from the Southern Adriatic (483 years, Barletta), avoiding data from Dalmatia and Croatia (Rovign, 262 and 254 years, respectively), because from a different geological context dominated by karst structures. We assume that, during the last climate cycle at least, the area where PRAD1-2 borehole was recovered was mostly influenced by the western Adriatic catchment area.
Five tephra layers determined by geochemical analysis occur within the investigated interval offering an independent cross check for the radiocarbon dates, and were included in the age model as additional control points. The five tephra were analyzed by Bourne et al. (2010) and the reader is referred this paper for further details about the tephra ages obtained independently from the $^{14}$C dates above described. Finally, three dated bio-events were used as additional control points (Piva et al. 2008) were used for the uppermost 2 m of the PRAD 1-2 borehole section and they include: 1) the Last Occurrence of the planktic foraminifer *Globorotalia inflata* at 6 cal. kyr BP, a biomarker recognized in the whole Adriatic basin and pre-dating the attainment of the maximum flooding during the Holocene (Trincardi et al., 1996); 2) the recognition of the time equivalent deposit of the Sapropel 1 in the eastern Mediterranean, a major oceanographic event centered at 8.5 cal. kyr according to the astronomical tuning by Lourens (2004); and 3) the paleoenvironmental change related to the end of the GS-1 event well recognized in the Adriatic basin and dated at 12 cal yr by Piva et al. (2008). In Table 4.2 all the available control points are listed. The control points obtained from of the radiocarbon ages are mid points.
<table>
<thead>
<tr>
<th>Sample top (m)</th>
<th>Age range (yr BP)</th>
<th>Source</th>
<th>Reference</th>
<th>Control points</th>
<th>SAR (cm/ka)</th>
</tr>
</thead>
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<tr>
<td>0</td>
<td>0</td>
<td>modern time</td>
<td>Piva et al. (2008) accepted</td>
<td>0</td>
<td>10</td>
</tr>
<tr>
<td>0.6</td>
<td>6000</td>
<td>LO G. inflata</td>
<td>Piva et al. (2008) accepted</td>
<td>6000</td>
<td>28</td>
</tr>
<tr>
<td>1.288</td>
<td>8500</td>
<td>Sapropel equivalent 1</td>
<td>Piva et al. (2008) accepted</td>
<td>8500</td>
<td>15</td>
</tr>
<tr>
<td>1.8</td>
<td>12000</td>
<td>Top GS-1</td>
<td>Piva et al. (2008) accepted</td>
<td>12000</td>
<td>18</td>
</tr>
<tr>
<td>2.18</td>
<td>14320 - 13900</td>
<td>Neapolitan Yellow Tuff</td>
<td>Bourne et al. (2010) accepted</td>
<td>14110</td>
<td>111</td>
</tr>
<tr>
<td>5.976</td>
<td>17226 - 17845</td>
<td>$^{14}$C</td>
<td>Piva et al. (2008) accepted</td>
<td>17540</td>
<td>108</td>
</tr>
<tr>
<td>7.8</td>
<td>18991 - 19569</td>
<td>$^{14}$C</td>
<td>Piva et al. (2008) mean</td>
<td>19275</td>
<td>420</td>
</tr>
<tr>
<td>7.84</td>
<td>19480 - 19050</td>
<td>Greenish/Verdoline</td>
<td>Bourne et al. (2010) mean</td>
<td>19498</td>
<td>43</td>
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<tr>
<td>8.8</td>
<td>19628 - 20065</td>
<td>$^{14}$C</td>
<td>this study accepted</td>
<td>21350</td>
<td>76</td>
</tr>
<tr>
<td>9.6</td>
<td>21024 - 21681</td>
<td>$^{14}$C</td>
<td>this study accepted</td>
<td>22528</td>
<td>72</td>
</tr>
<tr>
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<td>22376 - 22680</td>
<td>$^{14}$C</td>
<td>this study accepted</td>
<td>23780</td>
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<td>23471 - 24098</td>
<td>$^{14}$C</td>
<td>this study accepted</td>
<td></td>
<td></td>
</tr>
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<td>12.48</td>
<td>25602 - 26076</td>
<td>$^{14}$C</td>
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<td>12.78</td>
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<td>$^{14}$C</td>
<td>Piva et al. (2008) accepted</td>
<td>24725</td>
<td>25</td>
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<tr>
<td>13.32</td>
<td>28255 - 26302</td>
<td>VRa</td>
<td>Matthews et al. (2015) mean</td>
<td>27200</td>
<td>27</td>
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<tr>
<td>13.4</td>
<td>26741 - 27505</td>
<td>$^{14}$C</td>
<td>Piva et al. (2008) mean</td>
<td>32350</td>
<td>15</td>
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<tr>
<td>14.8</td>
<td>31536 - 33163</td>
<td>$^{14}$C</td>
<td>Piva et al. (2008) accepted</td>
<td>33300</td>
<td>26</td>
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<tr>
<td>14.94</td>
<td>33965 - 32630</td>
<td>Codola (base)</td>
<td>Matthews et al. (2015) accepted</td>
<td>39500</td>
<td></td>
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<tr>
<td>16.53</td>
<td>39390 - 39170</td>
<td>Campanian Ignimbrite</td>
<td>Bourne et al. (2010) accepted</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Table 4.2: the age model of the study encompassing the $^{14}$C dates on benthic species and tephra layers.

The radiocarbon dating at 12.48 m has been rejected as producing an inversion with the underlying age. The higher value may reflect interval-time with river input of older carbon-rich sediments. In two cases the age provided by $^{14}$C has shown an age overlapping with the age of a tephra positioned very close (within 4 to 8 cm), and therefore, to avoid possible distortions due to a too short integrating time interval, depth and age of the control points were calculated by averaging the values of the levels above and below. The age of the flooding surfaces has been calculated by linear interpolation between two successive control points, assuming constant the sediment accumulation rates between them. Finally, the sediment accumulation rates (SAR) have been calculated for each discrete interval.
4.2.2) Short-lived clinothem from the late Pleistocene Po River sediment wedge: an example of sand bypass to the basin

The late Pleistocene clinothem 300-400 is bounded by surfaces 300 and 400 at its top and base, respectively. Surfaces 300 and 400 are characterized by high amplitude and continuous reflectors (Fig. 4.1). Clinothem 300-400 is characterized by high amplitude and chaotic reflectors in the topset that pass to high amplitude chaotic and dipping and to low amplitude discontinuous and dipping reflectors at the foreset sector (Fig. 4.1).

Figure 4.1. Top left: seismic line LSD22 and clinothem 300-400; note the base of the late-Pleistocene sediment wedge (black line); note the subaerial rollover points (red dots) and the inflection points (black dots) of the clinothem 300-400. Note the location of the borehole PRAD1-2. Top right: Sea level curve: note the time interval and the depth interval of the paleo-shoreline where clinothem 300-400 developed. Below the sea level curves, the geometric measures of the clinothem 300-400 are reported (the time-depth conversion considers a mean of 1600 m/s in the sediment). Bottom left: seismic facies map of clinothem 300-400 (HA=high amplitude; LA= low amplitude; Ch= chaotic; D=discontinuous; CtoD= continuous-to-discontinuous; Dip= dipping; R= reflectors); note the average position of the paleo-shoreline during the development of clinothem 300-400. Bottom central: isochronal map of surface 300 (base of clinothem 300-400); note the presence of anticline structures in the central part of basin. Bottom right: thickness map of clinothem 300-400; note the minimum value at the shelf sectors and the depocenters developed along the strike axis of the basin, where two main depocenters are observable.

Across its bottomset sector, clinothem 300-400 is characterized by high amplitude continuous-to-discontinuous reflectors (semi-continuous reflectors), while at the toe, the clinothem 300-400 shows low amplitude continuous reflectors (Fig 4.1).
The isochronal map of surface 300 shows a gently dipping shelf, a curvilinear shelf edge, a steeper slope and a pronounced uneven topography at the basin sector characterized by a regional-scale convex-down geometry interrupted by convex-up geometries denoting a buried anticline in the central sector of the basin (Fig. 4.1). The thickness map of clinothem 300-400 shows a minimum thickness in the topset region (0-10 ms) and a prevalent accumulation in the foreset sector with depocenters elongated on E-W direction. Two main depocenters are located at the western and eastern side of the anticline. The western depocenter shows its maximum thickness up to 210 ms, while the eastern depocenters attains 150 ms (Fig. 4.1).

Clinothem 300-400 shows a descending rollover point trajectory with depths of the rollover point at 155 m at surface 300 and a depth of 175 m at surface 400 (Fig. 4.1). The height of the clinothem 300-400, taken as the vertical distance between the rollover point and the inflection point at the toe of the foreset (i.e. the foreset-bottomset transition: Pirmez et al., 2008), changes from 245 m on surface 300 to 178 m on surface 400; concurrently the horizontal distances between the rollover points and the inflection points decrease from 16 to 11 km (Fig. 4.1). The maximum vertical thickness of clinothem 300-400, measured at the forest, is 150 m. Surfaces 300 and 400 are intercepted by the PRAD1-2 borehole at 10.5 and 9.5 m below sea floor (bsf), respectively. Based on the age model from the PRAD1-2 (Table 4.2) the surface 300 have an age of 22.5 cal. kyr BP and the surface 400 have an age of 21.1 cal. kyr BP.

During the deposition of clinothem 300-400 the sea level was between -98 and -122 m compared to modern and likely the feeding system was close to the MAD as suggested by the vicinity of the paleo-shoreline to the rollover points (from few tens of meters to ca. 15 km; Fig. 4.1) and the presence in the topset sector of chaotic reflectors that are interpreted as amalgamated channels (chaotic fill sensu Mitchum et al., 1977; (Fig. 4.1). The very low aggradation recorded in the topset sector coupled with descending trajectories of the rollover points confirms that clinothem 300-400 developed during a fall in the base level. The base level fall was governed by the global fall in the sea level (Lea et al., 2002, among others) coupled with sediment compaction (measured up to 15% in the last 400 yr in
modern Po delta, Teatini et al., 2015), whereas tectonic subsidence can be considered negligible for that interval of time (Maselli et al., 2010).

Clinothem 300-400 shows a depocenter developed along the strike of the clinoform with two main depocenters on the western and eastern sides of an anticline. This evidence and the map of the surface 300 suggest that during its evolution the clinothem 300-400 conveys the structural confinement of the MAD (Fig. 4.1). Moreover, the depocenter tends to remain constricted on the northern part of the basin suggesting deposition under the influence of wave field (Fig. 4.4). The depocenters are characterized by semi-continuous high amplitude reflectors that resemble seismic facies encountered in intra-slope basins and interpreted as turbidite lobes and sheet-like turbidite (e.g. Normark et al., 1993; Gervais et al., 2006; among others). This finding implies that during the evolution of clinothem 300-400 sand was capable to bypass the shelf edge forming gravity-flow deposits at the base of the slope.

Based on the age model of table 3.2, clinothem 300-400 developed in a very short time-window of ca. 1.4 cal. kyr BP. During its deposition clinothem 300-400 reached very high accumulation rates up to 11.6 cm/yr at the foreset sector, two order of magnitude higher than 0.85 cm/yr of the modern Po Pila prodelta, and one order of magnitude higher than 1.8 cm/yr of the modern Po River subaqueous clinoform (Cattaneo et al., 2007). The very high accumulation rate may be explained by the enlargement of the catchment area of the ancestral Po River (ca. 200,000 km$^2$ against the modern ca. 74,000 km$^2$) and may reflect climate variability and increasing sediment flux during retreat phases of Apennine and Alpine glaciers (Giraud, 2011; Florineth and Schluchter, 1998).
5) **Compound clinoforms**

Seaward of major deltas worldwide, modern continental margins exhibit extensive muddy clinoforms, up to several tens of meters thick (Korus and Fielding, 2015). These clinoforms typically have sediment accumulation rates exceeding 1.5 cm/yr and have been building since the mid Holocene (stratigraphically above the maximum flooding surface attained at ca. 7-5 kyr BP), when sea level reached approximately its present position (Stanley and Warne, 1994; Nittouer et al., 1996; Asioli et al., 1996; Yi et al., 2003; Walsh et al., 2004; Amorosi et al., 2005; Liu et al., 2009; Korus and Fielding, 2015). Furthermore, such modern deltaic systems are characterized by compound clinoforms where progradation mainly takes place in two distinct areas: the coastal plain delta with a subaerial topset, and the subaqueous clinoform, composed of three geometric elements: topset, foreset and bottomset (Fig. 5.1; Nittouer et al., 1996; Steckler, 1999). The subaqueous deltas exhibit thickness distributions that appear strongly asymmetric with respect to their parent deltas, as a function of fluvial input and basin hydrodynamics (Korus and Fielding, 2015); because the effect of oceanographic processes redistributes river-borne sediment predominantly along the shelf, such clinoforms extend hundreds of kilometers away from major deltas, as far as 1500 km as in the case of the Amazon River (Nittouer et al., 1996).

![Conceptual sketch for model of delta progradation and associated compound clinoform development](image)

Figure 5.1: conceptual sketch for model of delta progradation and associated compound clinoform development: the shoreline rollover point (red circles) and shelf-clinoform rollover point (blue circles) are critical boundaries on continental margins (modified after Swenson et al., 2005; Nittouer et al., 2007).

In the last few decades compound delta systems have been received many attentions from both academy and oil companies because they represent a peculiar paleo-environment archive and
potential reservoir of the continental margin. Worldwide examples of compound delta systems may be found in both open continental margin settings and in semi-enclosed basins (Fig. 5.2), where subaqueous clinoforms develop tens of meter thick sediment successions with kilometric alongshore distributions: e.g., Amazon, (Kuehl et al., 1986); Eel river (Goff et al., 1999; Wheatcroft et al., 1997); Ganges-Brahmaputra (Kuehl et al., 1997); Yellow river – Shandong subaqueous clinoform (Alexander et al., 1991); Fly river (Walsh et al., 2004); Po river – Gargano subaqueous delta (Cattaneo et al., 2003); Rhone delta (Fanget et al., 2014). While subaqueous and compound deltas are increasingly recognized on modern continental margins (stratigraphically above the maximum flooding surface), their identification in the ancient rock record is hampered due to their reduced thickness and low angle of progradation that is difficult to detect at outcrop scales (typically less than 1° for muddy successions).

Figure 5.2: location of modern, and ancient, clinoform sets, from Patruno et al. (2015).

Earlier sequence stratigraphic models predict, during an overall seal level rise, dominant deposition on alluvial and coastal plain areas and sediment starvation conditions toward the shelf and in the basin (Posamentier et al., 1988). Conversely, among late Pleistocene to Holocene examples, a
growing body of evidence suggests that thick shelf progradations may form during an overall sea level rise. Examples of this came from Mediterranean continental shelves where seaward prograding deltaic to shallow marine successions formed during the short-term Younger Dryas climatic spell (Rabineau et al., 1998; Cattaneo and Trincardi 1999; Hernandez-Molina et al., 2000; Lobo et al., 2001; Aksu et al., 2002; Labaune et al., 2005, Berné et al., 2007; Jouet et al., 2006; Sømme et al., 2011).

The aim of the next manuscript is to understand if oceanographic condition similar to the modern one was already active in the Adriatic Sea promoting the development of a compound progradation. The stratigraphic record presented here shows, for the first time, the documentation of a well-preserved compound delta formed offshore the Gargano Promontory, offering the opportunity to characterize both coastal and subaqueous deposits developed during the Younger Dryas. The main goal in studying the Gargano compound delta can be summarized in three points: 1. Propose the possibility of a compound delta forming in a rapidly developed TST record to be used to re-evaluate cases of possible compound deltas recognized in outcrop or seismic surveys from ancient geological records; 2. Document a case of formation of a compound delta that encompasses even a shorter time than in the modern “high-stand” best-documented cases worldwide; 3. Infer the balance between oceanographic redistribution of fine-grained sediment and short-term climate-driven supply fluctuations from mainland.
5.1) Manuscript III

“Anatomy of a compound delta from the post-glacial transgressive record in the Adriatic Sea”

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Anatomy of a compound delta from the post-glacial transgressive record in the Adriatic Sea

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A B S T R A C T

On the Mediterranean continental shelves the post-glacial transgressive succession is a complex picture composed of seaward progradations, related to sea level stillstands and/or increased sediment supply to the coasts, and minor flooding surfaces, associated with phases of enhanced rates of sea level rise. Among Late Pleistocene examples, major mid-shelf progradations have been related to the short-term climatic reversal of the Younger Dryas event, a period during which the combination of increased sediment supply from rivers and reduced rates of sea level rise promoted the formation of progradations up to tens-meter thick. While the documentation of coastal and subaqueous progradations recording the Younger Dryas interval is widely reported in literature, the model of compound progradation within transgressive deposits has not yet been proposed. Here we present the documentation of a deltaic system where both delta front sands and related fine-grained subaqueous progradations (prodeltaic to shallow marine) have been preserved. The Paleo Gargano Compound Delta (PGCD) formed offshore the modern Gargano Promontory (southern Adriatic Sea), and is composed of a coastal coarse-grained delta of reduced thickness and a muddy subaqueous clinoform, up to 30 m thick. The PGCD, probably the first worldwide documentation of a compound delta within the transgressive record, provides the opportunity to investigate the processes controlling the formation of a compound delta system during an overall sea level rise and the factors that allowed its preservation. The finding of the PGCD provides the opportunity for a comparison with modern worldwide compound systems.

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1. Introduction

Modern and ancient river deltas represent one of the most intriguing sedimentary archives in the geological record, as their internal geometry and evolution reflects the interplay between river dynamics, sediment availability, grain size distribution and the oceanographic regime of the receiving basin (Orton and Reading, 1993; Cross et al., 1993; Trincardi et al., 2004; Slingerland et al., 2008). After the first process-driven classification of Wrigth and Coleman (1972), and Galloway (1975), another important step in understanding delta dynamics was introduced with the concept of compound deltas (Swenson et al., 2005), i.e., deltas composed by shallow-water progradations that are genetically related to deeper subaqueous clinothems. The transition in grain-size mostly related to the oceanographic regime of the basin, is also reflected by the overall geometry, characterized by two main roll-over points (i.e., breaking in slope at the topset/foreset transition; Swenson et al., 2005; Cattaneo et al., 2007; Walsh and Nittrover, 2009). The depth of the subaqueous rollover point is assumed to reflect the seaward limit beyond which wave-current shear stress decreases allowing sediment deposition (e.g., Nittrover et al., 1986; Kuehl et al., 1986; Alexander et al., 1991; Pirmez et al., 1998; Walsh et al., 2004).

Along the western Adriatic basin, the modern highstand deposits are organized in shelf-wide progradations that locally show variations in the stratal geometry: from North to South on bidimensional profiles it is possible to identify: 1) a single rollover point, where the delta front and the prodelta are “attached” (Fig. 1 section A–A'); 2) two rollover points (coastal and subaqueous rollover point), where the delta front and the prodelta are separated, and their distance is governed by the oceanographic regime (Fig. 1 section B–B', the 2D compound delta); 3) a single subaqueous rollover point where the subaqueous delta is disconnected from the delta front (Fig. 1 section C–C', purely subaqueous delta). In a section parallel to the modern shoreline from the modern Po River delta down to the Gargano Promontory, the Adriatic subaqueous clinoform has been interpreted as a “distorted” compound system (Fig. 1 section D–D'; Cattaneo et al., 2007).

The understanding of the physiographic parameters of the compound system is crucial to explaining the main processes and physical
Steckler et al. (1999) noticed that a subaqueous clinoform may prograde simultaneously with the coastal delta system depending on the oceanographic regime of the receiving basin; Swenson et al. (2005) highlighted that such genetically-connected clinoforms (coastal and subaqueous) may be geometrically disconnected depending on fair-weather and storm wave bases (10/15 m to 30 m water depth, respectively; Cattaneo et al., 2003) exhibiting a dominantly along-shore thickness distribution and a typical convex seaward morphology (Cattaneo et al., 2003). Examples of compound delta systems may be found in both open ocean settings and enclosed basins: e.g., Amazon River, (Nittrouer et al., 1996); Ganges–Brahmaputra River (Goodbred and Kuehl, 2000); Yellow River — Shandong subaqueous clinoform (Liu et al., 2004); and Po River — Gargano subaqueous delta (Cattaneo et al., 2003). Furthermore,
many authors focused on the subaqueous clinoforms that constitute tens of meter thick sediment successions with kilometric alongshore distributions far from the main Rivers: e.g., Ganges–Brahmaputra Rivers subaqueous delta (Kuehl et al., 1997; Palamenghi et al., 2011); Fly, Kikori and Purari Rivers (Walsh et al., 2004); Yellow River subaqueous delta (Yang and Liu, 2007); Mekong River subaqueous delta (Xue et al., 2010); Rhone River subaqueous delta (Fanget et al., 2014).

While compound and subaqueous deltas are increasingly recognized on modern continental margins, their identification in ancient sedimentary records is more difficult, mainly for two reasons: the difficulty in pinpointing the position of the feeder system and the poor geometric resolution in depicting the low angle of muddy subaqueous progradations (Mellere et al., 2002; Morris et al., 2006; Hampson, 2010).

The aim of this paper is to document a compound delta system (the Paleo Gargano Compound Delta, PGCD), formed during the post-glacial sea level rise, offshore the Gargano Promontory in the South Adriatic continental shelf. This study provides the opportunity to investigate the conditions that led to the formation and preservation of a complex deltaic system during a period of overall sea level rise and to develop new concepts for the study of ancient sedimentary archives. Among Late Pleistocene to Holocene examples, a growing body of evidence suggests that mid-shelf progradational deposits of variable thickness developed during the overall last post-glacial sea level rise (Fig. 2; Cattaneo and Trincardi, 1999; Hernández-Molina et al., 1994; Aksu et al., 2002; Labaune et al., 2005; Berné et al., 2007; Maselli et al., 2011; Semme et al., 2011) although no documentation of compound deltas has been proposed so far in these contexts.

2. Background

2.1. Geological setting

The Gargano Promontory peaks at 1050 m above sea level and has constituted a morphologic relief since the Middle Pliocene (Bertotti et al., 1999). The outcropping stratigraphic succession consists of Mesozoic carbonate rocks and includes four main lithostratigraphic units: the Maiolica Formation (Fm), the Marna a Fucoidi Fm, the Scaglia Fm and the Peschici Fm, characterized by thin- to thick-bedded lime mudstone, with Calcarenite and Brecia intervals (Eberli et al., 1993). The same lithology extends offshore the Gargano Promontory to form the Gargano Structural High (GSH) of the study area: a shallow plateau in ca. 50 m of modern water depth with steep flanks to the North and to the East (Fig. 1). The GSH represents an area of no or reduced sediment deposition during the modern highstand (Fig. 1; Cattaneo et al., 2003), where carbonate related karst features are exposed on the sea floor (Taviani et al., 2012).

2.2. Modern hydrology of the Adriatic Sea and sedimentation patterns

The oceanographic circulation of the Adriatic Sea is dominated by three main elements (Artegiani et al., 1997a,b): 1- a superficial cyclonic gyre with a component that flows parallel to the western Adriatic shoreline, mainly generated by the wind pattern; 2- the Levantine Intermediate Water (LIW), a salty water that forms in the Eastern Mediterranean Sea (Levantine Basin) and intrudes in the Adriatic basin flowing at depths of 200–600 m (Lascaratos, 1993); and 3- the North Adriatic Dense Water (NADW), that forms in the northern Adriatic through winter cooling (Vilibich and Supich, 2005; Benetazzo et al., 2014) and then, once a density threshold is reached, flows southward until cascading toward the deep southern Adriatic Basin (Trincardi et al., 2007; Canals et al., 2009). The overall thermohaline circulation, under the Coriolis apparent force, runs along the Italian coast constraining the main sediment flux to deposit in a prism parallel to the Apennine coast (Fig. 1; Correggiari et al., 2001; Cattaneo et al., 2003).

2.3. Stratigraphic setting of Adriatic TST deposits

In the Adriatic basin the Late Pleistocene–Holocene transgressive units comprise backstepping barrier lagoon deposits with large reworked sand dunes, in the northern low-gradient shelf (Trincardi et al., 1994; Correggiari et al., 1996), and subaqueous progradational deposits, in the western Adriatic shelf and in the Mid Adriatic Deep (MAD, Trincardi et al., 1996; Cattaneo and Trincardi, 1999; Maselli et al., 2011), where continuous chronological controls are available (Asioli, 1996; Blockley et al., 2004; Lowe et al., 2007). The transgressive systems tract (TST) is floored by an unconformity of regional extent (lowstand unconformity ES1) and topped by the maximum flooding surface (mfs).

On the central Adriatic shelf, the TST unit records the impact of sea level and sediment supply fluctuations on sub-millennial scales resulting in a tripartite TST (Cattaneo and Trincardi, 1999; Maselli et al., 2011). The lower and upper TST units (lTST and uTST unit, respectively) record an abrupt landward shift of the shoreline, while the middle unit (mTST unit) is prograding seaward and represents a regressive sedimentary body within the TST (Cattaneo and Trincardi, 1999). The three TST units are separated by two prominent, and extensively erosional, surfaces (S1 and S2 surface, Cattaneo and Trincardi, 1999;
Maselli et al., 2011; see Fig. 3 and Table 1 for a summary of the main stratigraphic surfaces and depositional units within the Late Pleistocene-Holocene TST record). In particular, the mTST unit is characterized by two sub-units (mTST-1 and mTST-2 sub-units) separated by an erosional surface (Si) that, possibly, records a minor sea level fall during the Younger Dryas interval (Maselli et al., 2011). The mTST-1 unit records the Bölling–Allerød interval, while the mTST-2 unit progrades during the Younger Dryas interval (Maselli et al., 2011).

From the central Adriatic to the southern Adriatic shelf the lTST and mTST units are organized in depocenters which reflect the interaction between sediment distribution, oceanographic processes and seafloor morphology. In particular, the Gargano Promontory is characterized by an articulated slope morphology and approaching the GSH the low TST and the mTST-1 units tend to diminishing in thickness and disappear, locally.

3. Data and methods

The database is composed by a dense network of high-resolution chirp-sonar profiles (acquired with a 16-transducer hull-mounted Benthos sound source), and sediment cores (both piston and gravity corers) acquired during the last decades by ISMAR-CNR with R/V URANIA (Cruises YD97, AMC99, CSS00, COS01, KS02 and SA03). The seismic profiles analyzed in the study area offshore Gargano Promontory sum to a total length of 1300 km over an area of 3200 km², with an average spacing between the lines of 2.5 km. A D-GPS allowed accurate positioning of the profiles and sediment cores.

Chirp profiles are imaged with a metric vertical scale assuming a 1500 m/s propagation sound velocity within sediment. By using a total length of 1300 km of seismic profiles, we constructed the structural map of the ES1 surface (the sequence boundary above which the transgressive record developed, Fig. 4), and the isopach maps of the main stratigraphic units composing the post-glacial stratigraphy offshore the GSH. The maps are stored in a GIS and projected in UTM with a WGS84 datum. In the lack of more precise velocity analyses on the sediments, volumes are quantified by assuming a constant 1500 m/s propagation sound velocity through superficial sediment.

Piston cores YD97-16 and KS02-330P were recovered on the GSH flank in 78.5 m and 79 m water depth using a 10 m barrel. The penetration of the corers was 7 m in the first site, and 11 m in the second, with a core recovery of 3.91 m and 6.6 m, for a total recovery of 70% in both cases. Both cores were sampled every 10 cm; core KS02-330P for grain size analysis and both cores for lithological and paleo-environmental reconstructions based on micro and macro faunal assemblages.

Calcimetry and grain size analyses were performed at Ifremer; the sampled fractions were disaggregated with a Retsch MM200 mixer mill at 17 cycles per second for 4 min. The CaCO₃ content was measured with an automatic pressure calcimeter (Dream Electronique model 2.1) while grain size analyses were made using a Coulter LS200 laser microgranulometer. Radiographies of the lowermost 3 m of core KS02-330P allow detection of bottom-current structures.

The correlation between cores and seismic profiles is performed after decompacting the total recovery versus the total penetration by assuming a linear compaction (the compaction is given by the ratio between the total penetration and the total core length). Correlation
<table>
<thead>
<tr>
<th>Seismic surface/age (yr BP)</th>
<th>Seismic unit and subunit</th>
<th>Corresponding sediment core depth (m bsl)</th>
<th>Description</th>
<th>Paleo-environment interpretation and sea level variation</th>
<th>Graphic sketch</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>HST</td>
<td>YD97-16 0-1.20</td>
<td>Prograding unit of the modern Po delta system and Gargano subaqueous muddy clinoform</td>
<td>Shelf-slope</td>
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<tr>
<td></td>
<td></td>
<td>KSQ2-330° 0-1.70</td>
<td></td>
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<tr>
<td>Mfs (ca. 5500)</td>
<td></td>
<td>YD97-16 1.20</td>
<td>Maximum flooding surface</td>
<td>Maximum adriatic sea ingestion</td>
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<tr>
<td></td>
<td></td>
<td>KSQ2-330° 1.70</td>
<td></td>
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<tr>
<td></td>
<td>uTST</td>
<td>YD97-16 1.20-2.20</td>
<td>Mud drape unit characterized by marine onlap terminations onto preexisting seafloor structures</td>
<td>Outershelf-upper slope</td>
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<tr>
<td></td>
<td></td>
<td>KSQ2-330° 1.70</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>S2 (ca. 11300)</td>
<td></td>
<td>YD97-16 2.20</td>
<td>Regional erosional surface</td>
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<tr>
<td></td>
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<td>KSQ2-330° 4.20</td>
<td></td>
<td></td>
<td></td>
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<tr>
<td></td>
<td>mTST-2</td>
<td>YD97-16 2.20-3.91</td>
<td>High-angle prograding unit</td>
<td>Inner-shelf (less than 30 m water depth)</td>
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<td></td>
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<td></td>
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<tr>
<td>S1 (12200-12800)</td>
<td></td>
<td>YD97-16 No recorded</td>
<td>Local erosional surface</td>
<td>Possible sea level fall during the YD interval</td>
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<tr>
<td></td>
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<td></td>
<td></td>
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<tr>
<td></td>
<td>mTST-1</td>
<td>YD97-16 No recorded</td>
<td>Very low-angle prograding unit</td>
<td>Inner-shelf (melting of Alpine and Apennine glaciers; increased sediment delivery)</td>
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<tr>
<td></td>
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<tr>
<td>S1 (ca. 14800)</td>
<td></td>
<td>YD97-16 No recorded</td>
<td>Regional erosional surface</td>
<td>MWP-IA</td>
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<tr>
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<td></td>
<td>KSQ2-330° No recorded</td>
<td></td>
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<tr>
<td></td>
<td>ITST</td>
<td>YD97-16 No recorded</td>
<td>Prograding unit with shingled reflectors</td>
<td>Early phases of the last sea level rise; influenced by fresh water input</td>
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<tr>
<td></td>
<td></td>
<td>KSQ2-330° No recorded</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>ES1 (ca. 18000)</td>
<td></td>
<td>YD97-16 Not reached</td>
<td>Regional erosional surface</td>
<td>Lowstand subaerial exposure (Sequence Boundary)</td>
<td><img src="image" alt="ES1" /></td>
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<tr>
<td></td>
<td></td>
<td>KSQ2-330° Not reached</td>
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among YD97-16 and KS02-330P cores relied on wiggle matching of the magnetic susceptibility logs.

4. Results

4.1. The middle TST unit offshore the Gargano Promontory

Offshore Gargano Promontory a prominent erosional surface (ES1), characterized on seismic profiles by a continuous and high amplitude seismic reflector, formed during the last glacial sea level lowstand, when the shelf underwent subaerial exposure, as highlighted for the northern and central Adriatic shelf by Trincardi and Correggiari (2000) when the shelf underwent subaerial exposure, as highlighted for the seismic record by Ridente and Trincardi (2002). On the steep northern and central Adriatic shelf by Trincardi and Correggiari (2000) when the shelf underwent subaerial exposure, as highlighted for the seismic record by Ridente and Trincardi (2002).

The middle TST offshore the Gargano Promontory is characterized on seismic profiles by a continuous and high amplitude seismic reflector, formed during the last glacial sea level lowstand, when the shelf underwent subaerial exposure, as highlighted for the northern and central Adriatic shelf by Trincardi and Correggiari (2000) when the shelf underwent subaerial exposure, as highlighted for the seismic record by Ridente and Trincardi (2002). On the steep northern and central Adriatic shelf by Trincardi and Correggiari (2000) when the shelf underwent subaerial exposure, as highlighted for the seismic record by Ridente and Trincardi (2002).

In this setting the mTST unit is confined between surface ES1 (below) and surface S2 (above), the latter corresponding to a high amplitude reflector of regional extent (Fig. 5A, B). In the proximal area of GSH the coastal deposits of the mTST unit show seismic facies with opaque discontinuous reflectors characterized by local and small scale steep reflectors (Fig. 5B). The seismic pattern of the coastal deposits shows high-angle foresets with a pronounced costal rollover point at ca. 60 m bsl (below sea level; Fig. 5A, B). The transition area between coastal and subaqueous deposits of the mTST unit is characterized by discontinuous reflectors in part due to the presence of gas charged sediment (Fig. 5A, B). In the area seaward of the GSH the subaqueous deposits of the mTST unit are characterized by complex sigmoidal seismic pattern reflector (sensu Mitchum et al., 1977; Fig. 5A, B). Within this subaqueous progradational deposits the subaqueous rollover point is detected at an average water depth of 90 m bsl (Fig. 5A B, C). The subaqueous rollover point shows an ascending trajectory that may suggest the onset of a relative sea level rise (Fig. 5A B, Helland-Hansen and Hampson, 2009).

The mTST unit develops in the proximal area and seaward of GSH over an area of $5 \times 10^3$ km$^2$ and with a total volume of about 4.6 km$^3$. This unit reaches a maximum thickness of up to ca. 30 m close to the GSH, where a bidirectional progradation develops (Fig. 5C), and then becomes thinner and spread over a broader area further to the south (Fig. 6). The bidirectional downlap is accompanied by an asymmetric internal geometry with steeper foresets (up to 2.2°), and a reduced seaward progradation toward the NW and gentler foresets (0.7°), with a greater seaward progradation toward the SE. Furthermore, the direction of progradation within the clinochrones (sensu Slingerland et al., 2008), determined from apparent angles measured along perpendicular profiles, is parallel to the coast. Interestingly, the mTST unit partly developed in the area where no HST units develop and presents two features indicative of highly energetic environments: a karst zone submerging about 12,500 years ago (Taviani et al., 2012), and a wave-cut terrace lie in the upper slope of the GSH at a modern water depth of about −55 m bfs (detail in Fig. 7).

In the YD97-16 and KS-330P sediment cores the mTST unit, below 2.20 m downcore and below 4.20 m downcore, respectively, is composed by fine to medium poorly-sorted sand (Fig. 8). The benthic assemblage shows the occurrence of species tolerating a high content of organic matter, i.e., Stainforthia complanata, Melonis padanum, Bulimina marginata, and Globobulimina spinescens, in addition to the inner shelf benthic species of Ammonia and Elphidium genera and miliolids, possibly indicating enhanced dysoxic conditions at the sea floor. The mTST unit is characterized by very few planktonic foraminifers typical of cold climate conditions (e.g., Neogloboquadrina pachyderma, Globigerina bulloides, Turborotalita quinqueloba). The discontinuous occurrence of opportunistic species like Valvulineria complanata and Nonionella turgida, coupled with episodes of dysoxic conditions, could either highlight an open mud-belt setting, characterized by a substantial content of organic matter and by reductive conditions at the sea floor, or be a consequence of repeated depositional events linked to river floods. The overall assemblage suggests that deposition occurred in an inner shelf environment. X-ray analysis highlights the pronounced cross...

![Digital Elevation Model (DEM, SRTM 90 m, from http://srtm.csi.cgiar.org) and structural map of the ES1 erosional surface at the base of the post-glacial TST record (grid 100 × 100 m), with contours every 10 ms. The Gargano Structural High (GSH) represents the drowned part of the Gargano Promontory. The map shows also the location of the seismic lines and cores discussed in this paper.](https://example.com/image.png)
4.2. The upper TST unit offshore the Gargano Promontory

The uTST unit is confined by surface S2, at the base, and the maximum flooding surface, on top. In the central Adriatic Sea, the uTST unit is a mud drape characterized by marine-onlap terminations onto pre-existing morphological structures (Cattaneo and Trincardi, 1999; Maselli et al., 2011). Offshore the GSH, the uTST unit shows sub-parallel and low-angle continuous reflectors with a progradational stacking pattern (Fig. 5A, B, C and detail in Fig. 7). The maximum flooding surface at the top of the uTST unit constitutes a prominent and continuous regional surface above which the highstand system tract develops (Fig. 5A, C and detail in Fig. 7). The maximum flooding surface displays an erosional character in the area close to the GSH, where the uTST unit shows a toplap termination (Fig. 5A, C). This evidence is indicative of a strong interaction between coast parallel currents and the pre-existing topography, resulting in an area of no deposition and or submarine erosion of the uTST unit (Fig. 7).

The thickness distribution of the uTST defines distinct depocenters separated from each other (Fig. 7). In the north-eastern portion of the GSH an area of no deposition or submarine erosion is evidenced by
the lack of the uTST unit. Because of the presence of this bypass area the uTST unit shows a detached geometry separated by the edge of the GSH (Fig. 7). The uTST unit advances mainly in southward direction, where the GSH is less prominent, showing a depocenter elongated in a N–S direction reaching ca. 21 m of thickness (Fig. 7). At regional scale the uTST depocenter marks a landward migration of the depositional area with respect to the underlying mTST unit, indicating a possible jump of the relative sea level (Fig. 7).

In the YD97-16 and KS-330P sediment cores the uTST unit, between 1.20–2.20 m and 1.70–4.20 m, respectively, consists of fine to medium
poorly sorted sand (Fig. 8). The foraminifera assemblage of *Ammonia* spp., *Elphidium* spp., *Nonion* spp., agglutinated taxa and miliolids testified a shallow water depositional environment. The uTST unit is characterized by a benthic fauna of *S. complanata*, *M. padanum*, *B. marginata*, *Ammonia tepida*, *Ammonia perlucida*, *Ammonia papillosa*, *Ammonia beccarii*, *Elphidium decipiens* and Miliolidae. This faunal assemblage is typical of a inner-shelf environment with nearby sediment input.

### 4.3. Chronological framework of the mTST unit

On the Adriatic margin, seismic stratigraphic correlations supported by 14C dating and biostratigraphic reconstructions allowed the correlation of the main TST units, and bounding surfaces, at a basin scale (Fig. 3 and Table 1; Trincardi et al., 1994, 2011; Cattaneo and Trincardi, 1999; Asioli, 1996; Asioli et al., 2001; Correggiari et al.,...)
2001; Cattaneo et al., 2003; Piva et al., 2008; Maselli et al., 2010). In the central Adriatic the mTST unit comprises two subunits each recording two intervals of sea level rise during the Bölling–Allerød and Younger Dryas events; these two mTST sub-units are separated by the Si erosion-al surface that developed during a minor sea level fall within the Younger Dryas cold spell (Maselli et al., 2011).

In the south Adriatic, offshore the Gargano Promontory, the ES1, S1 and Si surfaces coincide and the lower TST and mTST-1 sub-units are not recognized (Fig. 5). This may be related to a total cannibalization of the lower transgressive deposit during the Younger Dryas sea level fall proposed by Maselli et al. (2011) or a reduced deposition during the first phases of the post-glacial sea level rise, that may suggest that only the upper mTST unit proposed by Maselli et al. (2011), formed off-shore the Gargano Promontory. As suggested by several authors, transgression do not necessarily coincide in time along every part of a basin margin (Helland-Hansen and Gjelberg, 1994; Posamentier and Allen, 2001; Cattaneo et al., 2003; Piva et al., 2008; Maselli et al., 2010).

Fig. 9. Conceptual scheme for the PGCD showing the interplay between coastal sediment flux and alongshore sediment dispersal. The paleo-shoreline rims the GSH flank where down-drift current deflection leads to an asymmetrical compound systems characterized by a coastal and subaqueous rollover point at paleo-water depth of 3 and 28 m (red and blue dots, respectively). Inset: schematic compound delta modified after Nittouer et al. (1996), and Swenson et al. (2005).

Fig. 10. Correlation between mTST unit and sea level curve (modified after Fairbanks, 1989; Camoin et al., 2004; Liu and Milliman, 2004; Maselli et al., 2011). Average depths of seismic facies interpreted as possible wave-cut terraces (black circle), coastal rollover point (red dot) and subaqueous rollover point (blue dot) record a progradation phase during the Younger Dryas interval. Note that the coastal rollover point and wave-cut terrace occurred at average depths that implies their formation after the minor sea level fall proposed by Maselli et al., 2011, and during a phase of possible still stand in the sea level curve of Liu and Milliman, 2004. The water column between the coastal and the subaqueous rollover point was about of 25 m deep.
Seaward the GSH the mTST unit shows its maximum stratigraphic variability and complexity, with a well-preserved sandy coastal deltaic deposit spatially-connected and genetically related to a progradation wedge deeper on the shelf (Fig. 5). By analogy of modern compound systems, we propose that coarse-grained deposits formed at the shoreline, while fine-grained sediments accumulate in deeper shelf environments (typically in water depths of 25–30 m or deeper; Fig. 9). Given that the coastal rollover point of this compound delta is in a modern water depth of ca. 60 m, we can approximate its age to be ca. 11.8–12.6 kyr BP (Fig. 10, PGCD), by correlating the depth of the subaerial delta to global sea level curves (Fairbanks, 1989; Camoin et al., 2004; Liu and Milliman, 2004; Maselli et al., 2011), assuming that the vertical movements of the substrate in the area are negligible compared with the eustatic component.

As demonstrated for other Mediterranean margins, changes in precipitation rates and vegetation cover related to the Younger Dryas event may have resulted in enhanced sediment production in the catchments and/or enhanced rates of sediment export to the sea that favored the formation of thick progradational deposits (Cattaneo and Trincardi, 1999; Labaune et al., 2005; Berné et al., 2007; Sømme et al., 2011). Within this short time interval (ca. 800 years), the mTST subaqueous clinothene offshore the Gargano Promontory recorded a short phase of higher sediment accumulation rates (0.5 cm yr$^{-1}$, with a maximum of ca. 3 cm yr$^{-1}$ in the area where bidirectional downlap develops; Fig. 5C). The increased sediment supply in a relatively small interval of time can be explained by referring to the climatic reversal of the Younger Dryas event that promoted increased rates of sediment discharge from the rivers draining the high altitude areas of the Gargano Promontory that favored the development of compound delta (Fig. 9).
particular the subaqueous counterpart consists of basinward downlapping strata, indicating, for some time, that the sediment supply was able to counteract with the new accommodation space created by the ongoing sea level rise. Moreover, a greater sediment accumulation rate is also favored by focused deposition against the GSH: in this area the thickness of mTST unit reached ca. 30 m. Similar examples of focused sedimentation in depocenters down-current of obstacles like the GSH come from the modern subaqueous delta south of the Shandong Peninsula (Yang and Liu, 2007; Liu et al., 2009). These values are comparable with the sediment quantified accumulation rates recorded in modern subaqueous deltas worldwide (Fig. 11: at least 5 cm yr$^{-1}$ in the Ganges Brahmaputra foreset area, Michels et al., 1998; 7 cm yr$^{-1}$ in the Indus clinoform foreset, Giosan et al., 2006; up to 10 cm yr$^{-1}$ in the Amazon foreset area, Nittouer et al., 1986).

5. Discussion

5.1. Compound delta progradation and oceanographic regime

Seminal papers in seismic stratigraphy suggested that the presence of a rollover point at the topset-foreset transition may occur in marine environments and thus does not necessarily represent the shoreline.

Fig. 13. Top (A): comparison of the thickness distributions of the mTST (left) and the modern HST deposits (right). The middle TST unit is characterized by main depocenters reflecting local deltaic entry points and along-shore sediment redistribution. Right: maps of the HST (thickness more than 6 ms, modified from Cattaneo and Trincardi, 1999: Cattaneo et al., 2007). Compared to the mTST, the HST shows a more continuous depocenter from the Po delta to the area south of the Gargano Promontory (see also Fig. 1). Bottom (B): conceptual scheme of clinoform progradation of river-dominated “Gilbert-type” delta against subaqueous “advective-type” delta. In the first case (left side) the direction of progradation is parallel to the sediment transport because fluvial energy prevails on along-shore transport. In the second case (right side), the direction of progradation is perpendicular to the sediment transport, as the oceanographic regime dominates on river processes.
(Mitchum et al., 1977; Vail et al., 1977; Pratson et al., 2007), and that the rollover point may form at different water depths with several trajectories for a given trend of relative sea level rise, depending on sediment supply rates and oceanographic regime. Following this concept, Steckler et al. (1999) underlined that the ratio between sediment supply and accommodation space plays a fundamental control on the distance between the shoreline and its time equivalent subaqueous rollover point and assumed this distance to be at a minimum during intervals of sea level lowstand. While in these models the oceanographic regime is scarcely considered, Swenson et al. (2005) highlighted the importance of the wave and current fields as the main factor affecting the lateral separation between the coastline and the subaqueous rollover point. The extent of this separation depends on the shear stress on the seafloor that prevents the deposition in the foreset region, as pointed out by Pirmez et al. (1998). In Cattaneo et al. (2007), the compound delta model proposed by Swenson et al. (2005), was modified into a “distorted geometry” where the energetic geostrophic circulation limits the accommodation of sediment in the clinoform bottomset; in this view the seaward progradation is not limited by the lack of sediment but by increased energetic conditions.

The concept of compound delta can be applied also to the mTST unit offshore the Gargano Promontory, where a coastal sandy deposit is genetically linked to a distal subaqueous progradation (Figs. 9, 12). The difference in water depth of ca. 25 m between the coeval coastal and subaqueous rollover points is similar to the distance found in the modern Adriatic compound delta (ca. 25–30 m; Cattaneo et al., 2007). The two systems show comparable directions of progradation, with the two subaqueous clinothemes elongated parallel to each other and to their shorelines (Fig. 13A). This evidence suggests that during the deposition of the PGCD the oceanographic regime was characterized by a wave-current field similar to the modern one, and the interaction between wave energy and along-shore currents governed sediment partitioning between the subaerial and its subaqueous counterpart.

5.2. The main factors favored the PGCD preservation

The variability of transgressive deposits, driven by numerous factors influencing the shoreline migration, may be high for coeval deposits over relatively small distances within the same sedimentary basin (Heward, 1981). Minor changes in one controlling factor or a combination of the controlling factors may modify the preservation of transgressive deposits. The preservation of transgressive deposits is strongly influenced by a number of factors such as the depth of erosion, the wave-energy, the rate of relative sea level rise, the sediment supply, the along shore transport and the erosion resistance of bed material (Belknap and Kraft, 1981). Notwithstanding the preservation of coastal transgressive deposits is considered unlikely as a result of intense ravinement during shoreline transition, example of costal deposits within transgressive units on worldwide shelves are reported in the literature by several authors (e.g., Steel et al., 2000; Aksu et al., 2002; Salzmann et al., 2013; Gambier et al., 2014). Finally, progradational (regressive) coastal deposits are more likely to be fully preserved than those of transgressive coasts (Davis and Clifton, 1987); indeed if the sediment supply compensates or overwhelms the increasing accommodation space related to relative sea level rise an aggradational or progradational succession is expected, respectively (Curray and Moore, 1964; Galloway and Hobday, 1983).

The coastal counterpart of the PGCD is characterized by an along-coast sandy deposit connected to a muddy subaqueous clinoform (Fig. 9). We suggest that the preservation of this shallower unit may be related to a prevailing vertical component of submergence rather than the horizontal component of reworking deposits during the shoreline retreat. This is possible in contexts where the rates of sea level rise and sediment supply overwhelm the sediment dispersion rates of the currents (Davis and Clifton, 1987). This interpretation is in agreement with the “cliff overstep” model proposed by Zecchin et al. (2011) for steep shelves where relative high sea level rise connected to the Meltwater pulses promotes a rapid drowning of cliff portions with related rapid excursion of the shoreline, and with the documentation of the high sediment supply recorded on the Mediterranean shelves during the Younger Dryas interval (e.g., Labaune et al., 2005; Berné et al., 2007; Lericolais et al., 2010; Maselli et al., 2011; Samme et al., 2011). The further jump in sea level highlighted by the subsequent landward migration of the uTST unit depocenter (Fig. 7) leads to the burial of the mTST unit and, in turn, promotes its preservation. Furthermore, the internal reflector architecture and the external geometry of the uTST unit highlight that an along-shore current shaped the structure of the subaqueous clinothem progradation, as observed for the underlying mTST unit and the overlying HST unit. In this view, the PGCD highlights a peculiar case study where the interaction between a step-like sea level rise and a pre-existing morphology may have favored the preservation of coupled progradations in the area present down-current in respect to the structural high.

5.3. The PGCD in three dimensions

The PGCD and the modern Gargano subaqueous delta share a comparable internal geometry with the strike of the clinoform subparallel to the axis of the depocenters. This evidence implies that the transport of sediment occurs over significant distances parallel to the clinoform strike instead of perpendicular (or at high angle) to it. Fig. 13B shows an interesting implication of this concept: in a depocenter growing from a feeding point located upcurrent (the subaerial delta) the successive steps of deposition may lead to the formation of clinoforms that are normal to the direction of sediment transport and, therefore, have the youngest reflector located further down-current (Fig. 13B, left). In this case, the time lines are progressively younger ride off the subaerial delta in a fashion that is reminiscent of a subaerial Gilbert type delta. In the Adriatic basin, both modern and TST examples, instead, show time lines with a more radial pattern and display an increasing distance moving downdrift from the entry points. Examples of a similar progradation pattern are documented for the submarine delta of the Ganges–Brahmaputra (Michels et al., 1998), the modern clinoform of Gulf of Papua (Walsh et al., 2004) and the deglaciation deposits of the Gulf of Lyon (Labaune et al., 2005).

When a subaqueous delta is recognized on the continental shelf, it is genetically linked to a time equivalent deltaic or estuarine deposit, through which sediment is fed to the subaqueous counterpart (two examples come from the delta Po-Gargano, Cattaneo et al., 2007; and the Amapa coastal plain–Amazon compound systems, Nittouer et al., 1996). The morphologic profiles of compound systems are characterized by a couple of rollover points (i.e., coastal and subaqueous rollover points); this evidence is confirmed also in profile of worldwide rivers characterized by an estuarine outflow (e.g., Amazon and Columbia Rivers, Fig. 14A). Moreover, depending on the morphology of the basin, compound systems may be nourished by several point sources, captured by an alongshore current (e.g., Apennine and Po Rivers for the Po-Gargano deltaic systems, Cattaneo et al., 2003; Fly, Kiori and Purari Rivers, for the Gulf of Papua deltaic systems, Walsh et al., 2004). In compound systems where a well-developed costal delta grows, marine currents lead to an asymmetric progradation of both costal and subaqueous systems (Bhattacharya and Giosan, 2003; Cattaneo et al., 2003; Correggiari et al., 2005). Despite all the differences in the oceanographic regime among different systems, subaqueous deltas can always be viewed as the marine components of compound deltas which coastal counterpart may be located hundreds of kilometers away (e.g., Shan-dong subaqueous delta and subaerial Yellow River delta, Alexander et al., 1991; Yangtze subaqueous delta and tidal subaerial delta, Hori et al., 2002; Gulf of Papua subaqueous delta and Fly, Kiori and Purari subaerial delta, Slingerland et al., 2008; Indus mid-shelf subaqueous
delta and subaerial delta, Giosan et al., 2006; Ganges–Brahmaputra shallow subaqueous delta and subaerial delta, Palamenghi et al., 2011).

5.4. Sedimentation patterns of the PGCD

Modern compound deltas developed during the last ca. 5–8 kyr BP, and substantially after the post LGM sea level highstand was attained (e.g., Goodbred et al., 2003; Yi et al., 2003; Walsh et al., 2004; Cattaneo et al., 2007). In modern examples of compound systems, sediments carried by rivers are dispersed alongshore far from the river mouths, and accumulated in subaqueous clinoforms 20–40 m thick in water depth ranges of 30–90 (Liu et al., 2009). Fluvial sediment dispersed alongshore may be transported as far as 300 km (Mekong River, Ta et al., 2002), 700 km (Yellow River, Yang and Liu, 2007), and as much as 1500 km (Amazon River, Nittrouer et al., 1996). A particular case comes from the Yellow Sea, where the subaqueous clinoform is characterized by a bidirectional progradation (omega-shaped) due to the interplay between long-shore sediment transport and the physiographic constriction by the Shandong Peninsula (Yang and Liu, 2007).

The sediment accumulation rates calculated on the foreset region of the subaqueous delta is linearly related to the distance between the shoreline and the subaqueous rollover point, as highlighted for modern compound systems in Fig. 14B. Interestingly, the PGCD recorded slightly higher accumulation rates compared to the modern Gargano.
subaqueous delta (3 mm/yr and 1.28 mm/yr respectively), possibly because of a smaller distance between the source and the subaqueous clinothene during the Younger Dryas interval. The post-Younger Dryas sea level rise led to the drowning of the shelf and widening of the basin that triggered the formation of an alongshore current acting as a linear sediment source for the modern compound delta system. For this reason, in the modern Adriatic compound delta the distance between the feeder source and the subaqueous depocenter is larger (in order of 600 km).

During the deposition of the mTST unit, a sandy deposit about 12 m thick and its subaqueous counterpart developed asymmetrically over 50 km (Fig. 6), with the coastal and subaqueous rollover points separated by ca. 10 km, along a down-current dip section. As for modern analogs, this evidence may suggest the presence of an alongshore current that prevented the seaward progradation of the PGCD. In the up-drift side of the GSH the PGCD appears less developed. Conversely, in the thick and its subaqueous counterpart developed asymmetrically over an order of 600 km).

For this reason, in the modern Adriatic compound delta the distance between the feeder source and the subaqueous depocenter is larger (in order of magnitude greater than the modern value of Apennine rivers). The substantial absence of microfauna in the PGCD deposits offshore GSH suggests the influence of freshwater discharge, flowing from the continent and nourishing a delta NE of Vieste (Fig. 9). Considering an average sediment density of 2.5 g cm⁻³, the total sediment load of the river was in the order of less than 10 x 10⁶ t yr⁻¹, a value of one order of magnitude greater than the modern value of Apennine rivers (Frignani et al., 2005). These values may reflect the impact of a rejuvenated fluvial incision excavated during the Younger Dryas cold event documented on land (Amorosi and Milli, 2001).

6. Conclusions

In the southern Adriatic basin, mid-shelf deposits document the first case of a compound delta deposited within a period of overall sea level rise. Offshore the Gargano Promontory, the landward backstepping architecture of the Late Pleistocene to Holocene transgressive deposits was punctuated by the progradation of a compound delta system characterized by a sandy coast unit of reduced thickness and a subaqueous muddy clinothene up to 30 m thick. Seismic–stratigraphic correlations support the hypothesis that the Paleo Gargano Compound Delta (PGCD) formed in a short time window including the Younger Dryas cold spell, when slow rates of sea level rise and enhanced sediment production in the catchments promoted both coastal and subaqueous progradations. Likely, the increased rates of sea level rise during the post-Younger Dryas interval allowed the rapid drowning and preservation of the PGCD. More in general, our findings support the following conclusions:

• The formation of the PGCD within an overall transgressive record implies that the time required for the development of such deposit may take place during very short windows, likely in the order of centuries; this notion may be useful in interpreting ancient stratigraphic records where a much lower geochronological resolution may lead to assume or imply much slower rates of sediment accumulation.

• Depending on internal reflector geometries and overall architecture, the PGCD may be viewed as the transgressive, small-scale, analog of the modern highstand Adriatic compound delta, as both records reflect progradations influenced by similar alongshore sediment dispersal with sediment transport occurring dominantly along the strike of the clinoform.

The findings from the PGCD, in agreement with documentation from the literature (especially for the Amazon delta), suggest that subaqueous deltas are always connected to a coastal deposit, characterized by a coastal rollover point, that may be interpreted as the genetically linked subaerial counterpart of a compound delta system.

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6) Contourite deposits and clinoform growth

First introduced to define marine sediments deposited in the deep-sea by contour-parallel bottom thermohaline currents (Heezen and Hollister, 1964), the use of the term *contourite* has been broadened through time to embrace a larger spectrum of sediments that accumulate in a wide range of water depths by different types of bottom currents (Rebesco et al., 2014 for a review). Bottom current produce a variety of deposits and erosional features; the largest deposits are called “contourite drifts”. Based on their overall external morphology and taking into account different geological and oceanographic settings, drift systems can be classified in elongated drifts, mounded drifts, sheeted drifts, channel-related drifts, confined drifts, patch drifts, infill drifts, fault-controlled drifts, and mixed drift systems (Fig. 6.1; Rebesco et al., 2014).

![Figure 6.1: drift systems types and inferred bottom-current paths from Rebesco et al. (2014).](image-url)
The basal unconformity of a contourite drift is commonly revealed by a continuous high-amplitude reflector that may widen beyond the limits of the drift system, and which represents non-deposition or erosion produced by bottom-currents (Faugères et al., 1999); indeed, these bottom current are characterized by an average velocities well above threshold for deposition (Mulder et al., 2013). Within contourites, reflectors often display low-angle downlap onto the basal unconformity while tend to be characterized by toplap truncations against the upper unconformity (Stow et al., 2002). The overall internal seismic character of a drift is the typical pattern of continuous reflectors that tend to follow the gross drift morphology. This pattern reflects the long-lasting, semi-stable conditions that are the prerequisite for building up a large contourite drift, however, large temporary changes in current strength and sediment supply may occur during the lifetime of the contourite, causing shift between erosional and depositional environments (e.g. “out of phase drift” in Verdicchio and Trincardi, 2008; moat migration in Llave et al., 2011).

Contourite depositional systems contain an association of erosional and depositional sedimentary features, which are driven through a complex interaction of bottom currents and sediment supply, reflecting the oceanographic response on the variable atmospheric processes (Hernandez-Molina et al., 2014 among others). During the last decade, reflection-seismic investigations of contourite depositional systems have benefited greatly from the increased interest in deep-water area by the petroleum industry, which has lead to improved technologies and quality seismic data; for example a higher awareness of the contourite paradigm has revealed more small-scale examples along the ocean margins, as well as in shallow water environments and even within lakes (Fig. 6.2).
Contourites are well established and described from recent and sub-recent deposits although the same cannot be said for their identification in ancient sedimentary series exposed on land (Rebesco et al., 2014), because of the difficulty in depicting the low angle of fossil contourite strata and related moat, because of the lack of the three dimensions of the contourite system, and because of the lack of information regarding the paleoceanographic regime (Stow et al., 2008).

Although most of the large present-day contourite drifts have been initiated within the Neogene, they all experienced an intensification since the beginning of the Quaternary, whereas some responded even more vigorously with respect to the early to middle Pleistocene transition (Miramontes et al., 2015, among others); while stable isotope and sortable silt grain-size analysis highlight changes in the bottom current inflow and competence and variation in sediment supply on Milankovitch to millennial time-scale (Toucanne et al., 2007; Minto'o et al., 2015). Furthermore, a geotechnical approach highlighted the association between contourite deposits and marine landslide (Bryn et al., 2005; Masson et al., 2006; Cattaneo et al., 2014).

Regarding the south-western Adriatic margin, earlier publications highlighted the presence of contourite deposits, both in the basin and on the outer shelf, through the analysis of surficial
stratigraphy and sediment cores (Verdicchio et al., 2007; Verdicchio and Trincardi, 2008; Foglini et al., 2015). Based on their orientation, on their overall up-slope migration and based on their patchy distribution, contourite deposits in the Adriatic Sea have been interpreted as related to the interplay of the down-slope North Adriatic Dense Water, and the along-slope Levantine Intermediate Water (Verdicchio et al., 2007; Trincardi et al., 2007). The formation of these contourite deposits seem to be incentivized by an overall decreasing in bottom current velocity at the down-current sector of the Gargano Promontory (Martorelli et al., 2010). During time, these contourite deposits reflected changes in oceanographic setting and thermohaline circulation affecting the whole Mediterranean Sea; in particular these contourite deposits were less active during glacial periods probably because of the inhibition in the production of the North Adriatic Dense Water due to the subaerial exposure of the northern Adriatic shelf (Verdicchio and Trincardi, 2008), and the less intense production of the Levantine Intermediate Water in the Western Mediterranean compare to the present interglacial (Myers et al. 1998).

Until nowadays the timing of, and the factors leading to, the onset and subsequent evolution of contourite deposits in the Adriatic Sea remained unknown. The next chapter aims to reports the regional stratigraphic architecture of the Pliocene-Quaternary contourite depositional system, its spatio-temporal evolution and a long-term perspective on the variation of the bottom-current regime. This case history shows the onset of contourite deposits within the Pliocene succession where a sediment drift developed on the flank of a structural carbonate anticline, and may represent an example of source/seal hydrocarbon play. The findings from the south-western Adriatic margin not only shows that contourite drifts contributes to creation of the stratigraphic architecture of the margin, but also that contourites may form the bulk of shelfal progradational sequences. This finding highlights that far from a feeding system, lateral advection and current deposition may became the dominant mechanisms of progradation promoting the accretion of sediment drift within clinoforms, that in turn govern the final architectural motif of the margins.
6.1) Manuscript IV

“Pliocene–Quaternary contourite depositional system along the south-western Adriatic margin: changes in sedimentary stacking pattern and associated bottom currents”

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Pliocene–Quaternary contourite depositional system along the south-western Adriatic margin: changes in sedimentary stacking pattern and associated bottom currents

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Abstract The Pliocene–Quaternary history of the south-western Adriatic margin, represented by a complex contourite depositional system, records the palaeoceanography of the basin and the interactions between oceanographic processes and the uneven slope morphology that resulted from tectonic deformation. Three main stages can be recognized: (1) during the Pliocene, a giant sediment drift formed on the southern flank of the slope-transverse Gondola anticline that focused and accelerated the flow of slope-parallel bottom currents; (2) since the early to middle Pleistocene transition, a reorganization of bottom-current pathways led to a sharp change in the sedimentary architecture of the margin that became dominated by the growth of contourite deposits; (3) as of 350 ka, landward-migrating contourites on the outer shelf (less than 120 m water depth) reflect the presence of bottom currents also in shallow waters. This analysis of the sedimentary stacking pattern of the contourite depositional system that developed along the south-western Adriatic margin since the Pliocene enables disentangling the processes that controlled changes in bottom-current activity, demonstrating that bottom-current deposits constitute the bulk of depositional sequences at the Milankovitch timescale.

Introduction

The importance of contourite deposits as a key component of deep ocean environments was recognized since the seminal papers of Heezen and Hollister (1964) and Heezen et al. (1966). Over the last decades, contourite deposits have increasingly been identified in more shallow-water settings, from the base of the slope to the outer shelf (e.g. Fulthorpe and Carter 1991; Howe 1996; Stow et al. 1998; Verdicchio et al. 2007; Rebesco and Camerlenghi 2008; Van Rooij et al. 2010; Rebesco et al. 2014; Hanebuth et al. 2015). Viana et al. (1998) introduced the term "shallow-water contourite" to describe shelfal bottom-current deposits (water depths of 50 to 300 m) in contexts where the impact of waves and tides is negligible compared to a dominant contour-parallel geostrophic circulation.

In the Mediterranean Sea and particularly along its gateways, several examples of contourite deposits have been identified where long-lasting oceanographic regimes (maintained over timescales of 10^6 years) are constrained to preferential pathways. Prime examples are the middle slope of the Gulf of Cadiz (Hernández-Molina et al. 2006; Brackenridge et al. 2013), the Mediterranean Gibraltar gateway (Hernández-Molina et al. 2014), the outer shelf off south-western Mallorca (Vandorpe et al. 2011), the Corsica channel (Roveri 2002; Cattaneo et al. 2014; Miramontes et al. 2014), the Sicily channel (Marani et al. 1993; Verdicchio and Trincardi 2008a; Martorelli et al. 2011), the Apulian continental margin (Savini and Corselli 2010), and the south-western Adriatic margin (Verdicchio and Trincardi 2008b; Martorelli et al. 2010).

On the south-western Adriatic margin (SAM), previous studies explored the complex seafloor morphology and surficial stratigraphy, and highlighted the presence of contourite deposits and sediment drifts in both proximal and distal
settings (Trincardi et al. 2007a; Verdicchio and Trincardi 2008b). The present study attempts to document the timing of, and the factors leading to, the onset and subsequent evolution of a long-lasting contourite depositional system on the SAM. Focusing on a sector of the basin where the interactions between oceanographic processes and seafloor morphology are maximal, this paper aims to establish (1) the regional stratigraphic architecture of the Pliocene–Quaternary contourite depositional system, (2) its spatiotemporal evolution and (3) a long-term perspective on the variation of the bottom-current regime and its impact on changes in the depositional style of the system throughout time.

**Geological setting**

The modern seafloor morphology of the SAM reflects the interaction between long-term tectonic deformation and eustatic oscillations that controlled repeated changes in bottom-current circulation and sediment dispersal. The main offshore tectonic feature is the south Gargano deformation belt (Argnani et al. 1993) that includes an asymmetric relief called the Gondola anticline (Fig. 1). This compressional structure affects the shelf, the shelf edge and the slope, where it passes into the Dauno Seamount. The Dauno Seamount, at a height of up to 250 m from the basin floor, rises from the central sector of the SAM.

The south-western Adriatic continental shelf is up to 80 km wide in the Gulf of Manfredonia and becomes narrower further to the south (Fig. 1; less than 30 km), with dip gradients reaching up to 4% at the shelf break. The SAM shows an uneven seafloor morphology due to the presence of multiple mass-transport deposits (buried and exposed), testifying recurrent sediment failure (Minisini et al. 2006; Dalla Valle et al. 2015), and depositional/erosional sedimentary structures, including contourites, sediment drifts, furrows and scours that record the impact of energetic bottom currents today exceeding 60 cm/s (Fig. 1; Verdicchio and Trincardi 2008b; Foglini et al. 2015).

**Stratigraphic setting**

The stratigraphic architecture of the Mesozoic–early Cainozoic sedimentary succession of the SAM has been investigated through seismic profiles and exploration boreholes (Fig. 2). Two main seismic horizons can be detected and correlated at a regional scale: the M and Q reflectors. The M reflector is associated to the Miocene erosional surface (Fig. 2; Argnani et al. 1993), and marks the boundary between the Miocene carbonate succession below and the siliciclastic Pliocene–Quaternary succession above. The Q reflector represents the boundary between the Pliocene and the Quaternary succession (Fig. 2).

The chronostratigraphic framework of the late Quaternary succession in the Adriatic Sea was defined by the borehole PRAD 1.2 in the central Adriatic basin (Piva et al. 2008a, b). In this area, as in the SAM, the upper four depositional sequences, each bounded by an erosional surface of regional extent formed during sea-level lowstand, have been correlated to 100,000 year Milankovitch cyclicity (Ridente and Trincardi 2002; Ridente et al. 2009; Maselli et al. 2010).

**Oceanographic setting**

The SAM is affected by distinct water masses that include the Levantine Intermediate Water (LIW) and the seasonally modulated North Adriatic Dense Water (NAdDW) flowing along and across the slope (Fig. 1; Artegiani et al. 1997a, b). Both the LIW and the NAdDW are characterized by annual and inter-annual fluctuations in density, flow strength and sediment transport capacity (Turchetto et al. 2007). The LIW forms in the Levantine Basin through evaporation during the summer and cooling during the winter (Lascaratos et al. 1999). This salty water mass (29.0 kg/m³) enters the southern Adriatic through the Otranto Strait and follows a cyclonic path, with a water depth interval of 200–700 m (Fig. 1; the so-called South Adriatic Gyre: e.g. Artegiani et al. 1997a, b; Mantziafou and Lascaratos 2008).

The NAdDW forms on the shallow northern Adriatic shelf and its density increases (29.8 kg/m³) through winter cooling and evaporation associated with local wind forcing (Bora events). This cold water mass moves southwards along the Italian margin, flowing around the main capes (Ancona Cape and the Gargano Promontory), then cascades obliquely across the south-western Adriatic slope along the steepest sector reaching below the depth range impacted by the contour-parallel LIW (Trincardi et al. 2007b; Canals et al. 2009). Due to its dynamical properties (buoyancy and kinetic energy), a portion of the NAdDW remains trapped on the shelf, flowing as a contour-parallel bottom current over several hundreds of kilometres (Fig. 1; Benetazzo et al. 2014; Bonaldo et al. 2015).

Both the contour-parallel LIW and the NAdDW interact with the seafloor morphology, including the Gondola structure, the Dauno Seamount and the Bari Canyon. This leads to the formation of both depositional and erosional features in water depths between ca. 100 and 1,200 m (Fig. 1; Verdicchio et al. 2007; Foglini et al. 2015).

**Materials and methods**

The SAM has been investigated using (1) a dense grid of low-resolution multichannel seismic lines, property of the Italian Ministry of the Development, tuned with exploration boreholes, which allowed the regional correlation of the Pliocene–
Fig. 1 Top Pathways of North Adriatic Dense Waters (NAdDW) and Levantine Intermediate Waters (LIW) in the Adriatic Sea. Bottom Multibeam bathymetry of the study area (modified from Foglini et al. 2015), showing the erosional and depositional sediment features of the SAM contourite system. Note the trace of the Gondola deformation belt with an E–W to NW–SE hinge line swivelling to SW–NE in its slope sector, the Dauno Seamount.
Quaternary succession and the construction of the structural map of the M surface (Fig. 2); (2) a high-resolution multibeam bathymetry of the study area and dense grid of chirp-sonar profiles acquired by ISMAR-CNR during the last 10 years (Trincardi et al. 2014); and (3) a set of high-resolution multichannel and chirp seismic lines newly acquired by ISMAR-CNR during the ADRIASEISMIC 2009 and INV AS 2012 surveys.

Multichannel seismic lines were acquired by means of a water gun Sercel-S15 source and a teledyne mini-streamer with 24 channels (80–500 Hz frequency content). Chirp-sonar profiles were gathered using a hull-mounted 16-transducer source with a sweep modulated 2 to 7 kHz outgoing signal.

Results

Seismic stratigraphy

The stratigraphic boundaries M, Q, and the seafloor define three main successions from bottom to top: Mesozoic–early Cainozoic succession, below the M reflector; Pliocene–Quaternary succession and the overlying Pliocene–Quaternary succession, Note the disappearance of Miocene deposits in the Gondola well drilled on top of the Gondola anticline (modified from Argnani et al. 2009).
succession and Quaternary succession, respectively below and above the Q reflector (Figs. 2 and 3). The M surface is highlighted by a very high-amplitude and high-continuity reflector and marks the top of the Mesozoic–early Cainozoic succession (Figs. 2 and 3). The isochronal map of the M reflector shows an uneven topography characterized by regional-scale convex-up and convex-down geometries with anticline and syncline structures in the slope-basin area that encompass the Gondola anticline (Fig. 2). The Gondola anticline displays a northward orientation and a NW–SE hinge line on the slope switching to a NE–SW orientation at the base of the slope (Dauno Seamount). Northwards and southwards of the Gondola anticline, two syncline structures form local topographic lows with bases deepening from 1.2 s to 2.1 s from the shelf edge towards the basin (Fig. 2).

The Q reflector is highlighted by subtle downlap terminations and locally cuts the underlying strata documenting intensified bottom-current erosion (Figs. 3 and 4a, b). This surface separates the Pliocene deposits from the Pleistocene–Holocene succession (Argnani et al. 1993, 2009; de Alteriis 1995; Billi et al. 2007). Within the Quaternary succession, the ES8 unconformity is characterized by a high-amplitude reflector that cuts the underlying strata and marks a sharp change in the stacking pattern of stratigraphic successions (Fig. 3).

**SAM Pliocene succession**

The Pliocene succession is bounded by the M and Q reflectors at the base and top respectively (Figs. 3, 4a, b). The configuration of the Pliocene succession is characterized by sub-parallel reflectors (Fig. 3). The lateral continuity of the reflectors within the Pliocene succession is interrupted by the growth of the Gondola anticline structure where seismic reflectors are characterized by truncations and onlap terminations (Figs. 3, 4a, b). The configuration and geometry of these seismic facies varies laterally. Southwards of the Gondola anticline, the Pliocene succession shows lateral low-amplitude and discontinuous reflectors to continuous and high-amplitude reflectors (Fig. 3). Locally, the Pliocene succession is characterized by truncated reflectors with a vertical displacement in the order of 10 ms, and a divergent reflector configuration. The seismic reflector stacking pattern changes vertically, highlighting the increasing presence of mounded reflector geometries in the upper part of the Pliocene succession (Fig. 4b). In this upper portion, the Pliocene succession can be subdivided into mounded-parallel reflector packages of 15 ms thickness and characterized by a convex-up reflector geometry, onlapping the southern flank of the Gondola anticline. Both the convex-up crest and convex-down low show an up-dip
lateral migration towards the Gondola anticline crest (Fig. 4b).

The Pliocene succession fills the topographic lows of the Mesozoic–early Cainozoic succession and shows a predominantly mounded external geometry. The Pliocene succession appears to reflect the structural configuration, resulting in lateral variations in thickness from a few milliseconds on the Gondola anticline flanks, up to 400 ms towards the northern syncline lows (Figs. 3 and 4a, b).

**SAM Quaternary succession**

The Quaternary succession is bounded at the base by the Q reflector and by the modern seafloor on top (Figs. 3 and 4a, b). On the shelf and upper slope, the late Quaternary succession is internally subdivided by eight erosional unconformities labelled ES8 to ES1 progressively up-section. In the basin, only the deepest three depositional sequences (8 to 6) are preserved (Fig. 4a, b).
Depositional sequence 8, between the Q reflector and the ES8 unconformity, is characterized by sub-parallel and semi-continuous reflectors that show a vertical change from low- to high-amplitude seismic facies (Fig. 4a, b). Locally, the lateral continuity of the reflectors is interrupted by the presence of the Gondola anticline (Fig. 4b).

The depositional sequences 7 and 6, bounded respectively by ES8–ES7 and ES7–ES6 unconformities, are similar to depositional sequence 8 in terms of seismic amplitude and internal reflector geometry, with the exception that they cannot be correlated basinwards beyond the Gondola anticline (Fig. 4b). Depositional sequence 5 is bounded by the ES6 and ES5 unconformities at the base and top respectively, and is characterized by a high-amplitude seismic facies on the outer shelf with continuous and sub-parallel reflectors (Fig. 4a, b). Landwards, the reflectors of depositional sequence 5 display toplap truncations, while basinwards they are truncated by the head scarp of the Gondola slide. The Gondola slide cuts the margin sequences down to the ES6 unconformity, above which the slide deposit is characterized by low-amplitude seismic facies and discontinuous to chaotic reflectors (the Gondola slide; Fig. 4a, b).

On the slope and in the basin, the Quaternary succession is characterized by an overall sigmoidal to oblique reflector configuration and includes elongated mounds (drifts A, B and C in Fig. 3). Regional stratigraphic correlation between drifts A, B and C with depositional sequences developed on the shelf was hampered by the lateral discontinuity of the reflectors (Figs. 3 and 4a, b). Drift A is characterized by high-amplitude seismic facies passing to low-amplitude seismic facies in the trough area (Fig. 3). This drift has upward-convex reflectors with downlap terminations on the northern flank and truncated terminations on the southern flank, suggesting decreasing deposition and locally enhanced erosion. Internally, drift A is characterized by repeated downlap terminations that help identify packages of mounded-parallel reflectors. The crests of the mounded reflector packages have axes verging towards a moat (Fig. 3). Drift A has a thickness of up to 300 ms constituting the northern flank of the Bari Canyon (Fig. 3).

Drift B is characterized by high-amplitude seismic facies deposits with sub-parallel reflectors of overall mound geometry (Figs. 3 and 5). Internally, this drift shows packages of discontinuous to chaotic reflectors interlayered with packages of continuous mounded reflectors (Fig. 5). Drift B is located on the southern flank of the Gondola anticline and reaches a maximum thickness of 150 ms (Fig. 5).

Drift C is characterized by high-amplitude seismic facies passing to low-amplitude seismic facies in the trough and toe areas (Fig. 3). This drift has mounded-parallel reflectors. Locally, the data reveal packages of undulating reflectors with wavelengths of up to 1.3 km and amplitudes of 50 ms, and with crests that show an upslope-verging axis. Drift C is located on the northern flank of the Gondola anticline and reaches a thickness of up to 200 ms (Fig. 3).

Unconformities ES4 to ES1 correlate with those recognized by Ridente et al. (2007) on the Apulian shelf and in the central Adriatic, and the depositional sequences 4 to 1 of the present work also coincide with those of Ridente et al. (2007). Depositional sequence 4 is bounded by the ES5 and ES4 unconformities at the base and top respectively (Fig. 4a, b). This depositional sequence is characterized by shingled continuous reflectors that are truncated basinwards against the modern edge scarp (Figs. 4a, b and 6a, b). There is a remarkable change in the stacking pattern from units 8–4 to units 3–1 resting above the pronounced erosional ES4 unconformity (Fig. 4b). The depositional sequences 3 to 1 are bounded by the ES4–ES3, ES3–ES2 and ES2–ES1 unconformities respectively (Figs. 4a, b and 6a, b). These depositional sequences are characterized by the presence of shingled, seaward-dipping reflectors and wavy reflectors locally truncated by the overlying unconformity (Fig. 6a, b). The wavy reflectors are characterized by a 100-m-scale wavelength and a 10-m-scale amplitude. The crests have landward-verging axes that document a consistent upslope migration of the sedimentary features. The reflectors dip and onlap consistently landwards compared to the incisions (towards W/SW; Fig. 6a, b).

Discussion

SAM contourite depositional system

Based on their seismic facies, internal stacking pattern, external drift morphology (Faugères et al. 1999; Rebesco and Stow 2001; Rebesco et al. 2014), as well as existing knowledge on the oceanographic setting and regional bathymetry (Trincardi et al. 2007b, 2014), the sedimentary bodies recognized along the SAM can be interpreted as contourite deposits. A large variety of erosional features are associated with contourite deposits, including large drift moats, scours and widespread discontinuities (Figs. 1, 3 and 6a, b). Contourite deposits and related erosional features have developed in the basin, along the slope and on the outer shelf, and include mounded drifts, separated drifts and shallow-water contourites (Figs. 3 and 6a, b).

The Pliocene succession is composed of a giant sediment drift, adjacent to a moat (Fig. 4b) that developed on the southern flank of the Gondola anticline at the modern water depth of 460–660 m (Fig. 7a). Comparable bottom-current deposits were recognized at the toe of structural highs in different continental margins, where an increase in shear stress and velocity is related to confinement of the flow against an obstacle (e.g. Van Rooij et al. 2010: Le Danois drift, Cantabrian margin; Martorelli et al. 2011: south-western seamount, Tunisian
margin). As testified by their external geometry and orientation, drifts A, B and C are interpreted as upslope-migrating sediment drifts. Similar Quaternary examples come from different settings where, due to topographic constrictions, bottom currents reached velocities sufficient to erode and redistribute fine-grained sediments (Faugères et al. 1999; Roveri 2002; Ercilla et al. 2002; Verdicchio and Trincardi 2008a; Van Rooij et al. 2010). Drift A developed on the slope and is part of the northern flank of the Bari Canyon where cascading dense shelf water currents are laterally confined (Trincardi...
et al. 2007b). Drifts B and C occur on both flanks of the Gondola anticline (Fig. 3); these separated drifts developed after the formation of the fault-generated relief of the Gondola anticline. Note the contrasting stratigraphic geometries within the Pliocene succession in the two sub-basins and their interaction with tectonics. On the outer shelf the internal architecture of the depositional sequences reveals the presence of sediment wave deposits (Fig. 6a, b). These display a pronounced landward migration pattern and are interpreted as contour sediment waves of 100 m wavelength.

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The complete system of contourite deposits and related erosional features of the SAM present characteristics similar to those of contourite depositional systems found worldwide on continental margins, and are the result of strong interactions between bottom-hugging current activity and irregular seafloor morphology. Key examples include the US Atlantic continental rise (Locker and Laine 1992), the Weddell Sea, Antarctica (Maldonado et al. 2005), the Argentina slope (Hernández-Molina et al. 2009), the Gulf of Cadiz (Llave et al. 2007) and the north Iberian margin (Van Rooij et al. 2010).

Fig. 7 Conceptual scheme of the Pliocene to Quaternary depositional history of the southwestern Adriatic margin in a section from the outer shelf to the basin across the Gondola anticline. a Sediment drift within the Pliocene succession related to the onset of a bottom current (orange arrow) flowing close to the southern flank of the Gondola anticline. Note the contrasting stratigraphic geometries within the Pliocene succession in the two sub-basins and their interaction with tectonics. b Change in the stacking pattern of Quaternary deposits above the ES8 unconformity (blue line). Note the growth of the contourite depositional system in a broader margin area, ascribed to the interplay of along-slope and down-slope currents (LIW and NADW respectively). c From the time of deposition of sequence 3, the bottom currents begin to affect a large portion of the outer shelf. Note that in this sector of the margin the bottom-current deposits constitute the bulk of the depositional sequences.

**Pre-contourite depositional phase**

The Pliocene succession immediately above the M surface shows a low-amplitude reflector deposit with a poorly pronounced mounded external geometry (Fig. 4b). This deposit records a pre-contourite depositional phase developed under weaker bottom-current activity. Similar deposits were recognized in the Algarve margin and interpreted as being associated with a pre-contourite sedimentation phase in the Early Pliocene succession (Llave et al. 2011), where the influence of down-slope sedimentary transport can not be excluded (Brackenridge et al. 2013).

**Onset of bottom-current deposits and their evolution**

Based on the available regional stratigraphic correlations, the onset of the bottom-current deposits along the SAM occurred during the Pliocene. The Pliocene sediment drift developed on the southern flank of the Gondola anticline and is characterized by packages of mounded-parallel reflectors that highlight the presence of contourite units (Fig. 4b). The stacking pattern of the contourite units with a typical thickness of 15 ms may suggest changes in bottom-current velocity and related...
The evidence of truncated and deformed reflectors in both sub-basins (on the northern flank of the Gondola anticline and beneath the Gondola slide; Fig. 3) confirms the tectonic activity during the Pleistocene and suggests a strong interaction between the Gondola slide and tectonic activity as highlighted by previous works (Minisini et al. 2006; Ridente et al. 2007). A close view of drift B reveals the presence of mass-wasting deposits interfingered with contourite units (Fig. 5). As interpreted in other settings, rapidly accumulated contourite drifts have often been associated with slumping phenomena due to the high water content, critical angle of deposition and the presence of weak layers (e.g. Laberg et al. 2003; Bryn et al. 2005; Kvalstad et al. 2005; Masson et al. 2006; Berndt et al. 2012; Ai et al. 2014; Cattaneo et al. 2014). In the case of drift B, moreover, the failure of the contourites may also be related to the activity of the Gondola anticline and related fault system, suggesting multi-event Pleistocene–Holocene tectonic activity. An extreme example is represented by drift C, which is almost entirely destroyed by the Gondola slide (Fig. 3).

These findings highlight an overall change of bottom-current activity through the Pliocene–Quaternary, with the development of multiple and perhaps more interchangeable current pathways impacting on a broader basin area (Fig. 7b). This current activity reflects a significant climatic reorganization after the early to middle Pleistocene transition when the 100,000 year cyclicity became dominant (Ruddiman et al. 1989). Furthermore, as pointed out by Selli (1965), an intensification of alpine glaciations occurred approximately 800 ka ago and records the onset of the so-called glacial Pleistocene in the Mediterranean region. On the SAM, the pacing of global Pleistocene climate oscillations is likely associated with changes in the strength and behaviour of thermohaline currents. These changes are highlighted by the progressive adjustment in the architectural pattern and by the changing growing style of the contourite and waveform deposits that developed in progressively shallower depths compared to the Pliocene drift deposit (Figs. 4b and 7b).

Onset of shallow-water contourites and their evolution

On the south-western Adriatic margin, the late Quaternary succession is composed of progradational deposits, each several tens of meters thick and separated by shelf-wide unconformities related to 100,000 year glacio-eustatic cycles (Ridente and Trincardi 2002; Ridente et al. 2009). The internal architecture of the three uppermost depositional sequences (D.S.1 to D.S.3, from top to bottom) shows, in particular, that shallow-water contourites constitute the bulk of each sequence, contributing significantly to the margin architecture and to the development of the south-western Adriatic contourite depositional system (Figs. 6a, b and 7c). The internal geometries of each depositional sequence, characterized by short-distance lateral variability in sedimentary architecture, may be related to changes in the strength and direction of bottom currents on the outer shelf in response to Quaternary climatic and oceanographic turnovers. As pointed out by several authors, sea-level changes at 100,000 year cyclicity governed the strength of the thermohaline circulation (e.g. Myers et al. 1998) and indirectly played a fundamental role in the development of shallow-water contourites (e.g. Viana et al. 1998; Verdicchio and Trincardi 2008b; Vandorpe et al. 2011).

In the SAM, in particular, the stratal pattern may reflect enhanced dense water formation in the northern Adriatic basin, and/or intervals of enhanced intrusion of the LIW towards the outer shelf with a lateral shift of the LIW–NAdDW...
interface. The shift of this interface towards shallower depths could be governed by eustatic cycles (Rogerson et al. 2012), which in turn affected the shift towards a shallower depth in the area where NAdDW is entrained with the LIW. Moreover, the internal complexity of each depositional sequence developed on the outer shelf may be related to changes in bottom-current strength and/or decrease in sediment supply. The presence of unconformities above each depositional sequence highlights the substantial reduction of accommodation space during relative sea-level fall, in turn leading to shallow-water erosion of contourite deposits accumulated during the preceding sea-level highstands (Fig. 6a, b). Moreover, these findings suggest the onset of current pathways similar to modern conditions, where the interplay of along-slope and down-slope (cascading) dense shelf water masses, impacting on a progressively broader margin area (extending to deeper and shallower depths), led to the development of sediment drifts with a patchy distribution (Fig. 7c).

Conclusions

The sequence stratigraphic analysis of bottom-current deposits along the Mediterranean margin and Mediterranean gateways helps identify the onset and evolution of sediment drifts, from which the associated bottom-water circulation can be inferred. These bottom-current deposits developed after the restoration of open marine conditions in the Mediterranean following the Messinian salinity crisis. On the south-western Adriatic margin, the formation of a contourite depositional system has been magnified in an area of complex slope morphology, where tectonic activity during the Pliocene–Quaternary has given rise to the slope-transverse Gondola anticline. This shows that:

- since the Pliocene a giant sediment drift migrated upslope, recording the ancestral LIW activity focused on the southern flank of the Gondola anticline;
- close to the early to middle Pleistocene transition, the impact of bottom currents spread to a broader slope area, leading to the generation of large-scale contourite deposits as well as upslope-migrating sediment waves and associated erosional moats;
- starting from ca. 350 ka (i.e. since the deposition of sequence 3), bottom currents affected for the first time a large portion of the outer shelf in shallow-water and more proximal settings, and bottom-current deposits constitute the bulk of the depositional sequences at Milankovitch timescales.

The progressive adjustment in the architectural pattern and the growing style of contourite deposits with a patchy distribution and different orientations seem to be related to the onset of a bottom-current pathway similar to the modern one that impacted on a broader and uneven seafloor morphology. It is possible that these bottom-current deposits reflect the activity of the Levantine Intermediate Water (today confined between 200 and 600 m but capable of intruding shallower shelf environments) and the onset of preferential pathways of the NADW that in part remains on the shelf without cascading obliquely to the slope.

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Conflict of interest

The authors declare that there is no conflict of interest with third parties.

References

7) Conclusion

This dissertation offers an improved understanding of the evolution of the Adriatic Basin during the Quaternary, in particular focusing on how changes in river hydrodynamics, accommodation (eustatism, tectonic activity and sediment compaction) and climate (sediment flux and oceanographic regime) controlled sediment transport and continental margin growth. To reach this goal, my Ph.D. work focused on four case histories along an hypothetical sediment routing system, from the modern Po River (chapter 3), down to the Adriatic continental shelf (chapters 4 and 5), and basin (chapter 6).

The first case history is a portion of the Po River used as a natural laboratory to quantify how the backwater effect created by a dam-induced reservoir controls river hydrodynamics, sediment transport and river morphodynamic evolution, without introducing assumptions on the effect of waves and tides. The data show that lateral migration rates strongly decrease across the backwater zone due to a reduction of flow velocity and bed shear stress; this hydrodynamic condition promoted the deposition of coarse-grained material, moving upstream the gravel-sand transition of the river. Further downstream, close to the dam, river bed sediments are progressively finer, up to clay size, while river bedforms become progressively wider due to a combination of decreased flow velocity and sediment grain size. The results obtained may be used for a better understanding of the morphodynamic evolution of rivers, deltas and estuaries.

In the second case history, from the central Adriatic shelf (chapter 4), the Po River lowstand wedge accumulated between 31.8 and 14.4 kyr B with a 350-m-thick succession during repeated phases of progradation, aggradation and degradation on millennial scale (1-3 kyr). Thanks to the extremely high spatial and temporal resolutions achieved, the findings from the Po River lowstand clinothems highlight that sand bypass toward the basin, expected principally during periods of sea level fall and lowstand, is not steady even on a very short-lived lowstand (ca. 15 kyr in duration): high-frequency base level oscillations may force the development of sandy deposits separated by flooding surfaces and relatively thick muddy clinothems promoting the compartmentalization of potential reservoirs.
In the third case history, from the area offshore Gargano Promontory (chapter 5), late Pleistocene to Holocene mid-shelf deposits document the first case of a compound delta developed within a short-time window during the post glacial sea level rise. The scientific literature on worldwide modern continental margin so far has documented the presence of these deposits only in highstand conditions, and so above the maximum flooding surface. Offshore the Gargano Promontory, a compound delta system (PGCD) is documented as part of a landward backstepping transgressive unit; this compound delta is characterized by a sandy coastal unit up to 10 m and by a genetically-linked subaqueous muddy clinothem up to 30-m-thick. The rapid formation of the PGCD (between ca. 11.8-12.6 kyr BP) implies that the maximum time required for the development of this kind of deltaic deposit is in the order of just few centuries. This notion may be useful when interpreting ancient stratigraphic records where a much lower chronological resolution available may lead to assume or imply much slower sediment accumulation rates. Moreover, the finding from the PGCD provides the opportunity for a comparison with modern compound systems and gives the possibility to extract information on the paleo-oceanographic regime. In this view, the PGCD may be viewed as the transgressive, smaller-scale, analogue of the modern highstand Adriatic compound delta, as both records reflect progradations influenced by similar southward-directed alongshore sediment dispersion with transport occurring dominantly along the strike of the clinoform with limited exploitation of accommodation in the topset region. The comparison of the external geometry of both systems, characterized by a vertical distance of ca. 25 m between the coastal and subaqueous rollovers, suggests that during the deposition of such compound progradations the oceanographic regime was characterized by a wave-current field similar to the modern one. Conversely to the modern Po-Adriatic compound delta that is characterized by a continuous depocenter, the PGCD and the overall distribution of the middle TST unit, highlight the presence of distinct depocenters along the Italian coast indicating that an along-shore sediment dispersal similar to the modern one was already active but either without reaching a competence similar to the modern one or not receiving enough sediment from the continent to allow the creation of a continuous depocenter along hundreds of kilometers as
in the case of the late-Holocene. Finally, the finding from the PGCD, in agreement with documentation from the literature, suggests that subaqueous deltas are always connected to a coastal deposit, characterized by a coastal rollover point that may be interpreted as the genetically-linked subaerial counterpart of a subaqueous progradation. The coastal deposit may constitute the geometrically and genetically linked coarser-grained reservoir of hydrocarbons migrating from the muddy and potentially organic-rich subaqueous counterpart. This concept may be useful when interpreting lower resolution seismic profiles where the spatial relation between coastal and subaqueous correlative deposits may remain more difficult to establish.

In the fourth case history, from the southern Adriatic Basin (chapter 6), the study of the southwestern Adriatic contourite depositional system developed since the Pliocene provides the opportunity to reconstruct the current pathway on a longer-term perspective. Three main phases of margin growth were recognized, each characterized by specific current pathways and, possibly, intensity: during the Pliocene a cyclonic circulation along a pattern similar to the modern Levantine Intermediate Water interacted with the paleo-morphology of the basin, characterized by two sub-basins northward and southward of the slope transverse Gondola anticline. The impact of the bottom currents was initially focused prevalently on the down-current zone, the southern sub-basin, where a sediment drift up to 230-m-thick developed; a successive phase of margin growth reflects a significant climatic reorganization during the early to middle Pleistocene transition, when bottom currents spread to invest a broader slope area, leading to the generation of large-scale contourite deposits as well as upslope-migrating sediment waves, resulting from the interplay of the Levantine Intermediate Water and the North Adriatic Dense Water. During the phase of margin growth that onset at ca. 350 kyr, bottom currents impacted on the outer shelf, likely in response to an excess in buoyancy (loss of salt content or increase in water temperature), and contourite deposits became a significant contribution to the bulk of the progradational sequences that record a 100 kyr cyclicity, promoting the shelf progradation by accretion of successive contourite drift. In conclusion, it seems that a dominant cyclonic Adriatic circulation pattern (related to a water mass like the modern Levantine Intermediate
Water) was well established by Pliocene time and during the climatic reorganization close to the early to middle Pleistocene transition. The sediment drift development was possible under this paleoceanographic regime because large volumes of sediment were introduced to the region (off-shelf marginal platform). This finding highlights that far from a feeding system, the accretion of sediment drifts within shelf-margin clinoforms may represent a significant contribution to the margin growth and in turn govern the final architectural motif of the margin. The contourite deposits of the south Adriatic margin developed during short-time intervals and with high sediment accumulation rates; this evidence, combined with the fine-grained lithology of the deposits and their very high water content, promoted the instability of the margin and the formation of repeated thin-skinned landslides, suggesting that contourite deposits contain potential weak layers or promote failure through rapid compaction and fluid escape processes. Sediment drifts located above drowned carbonate platforms, as in the case of the Pliocene sediment drift, possibly comprise an interesting hydrocarbon play. The sediment drifts, located on the flank of a structural relief, are well placed to intercept hydrocarbons migrating from underlying carbonate source beds. Furthermore, these deposits have depositional closures and may be capped by muds. If clean sands are present, drifts of this type have reservoir potential and may be worthwhile exploration targets in some basins.

In all of the three case histories focus on Adriatic depositional basins (chapter 4-6), the depositional history can be constrained by a very-high chronological framework that allows documentation of short-lived phases of extremely rapid deposition during the margin construction. A common depositional motif is the growth of clinothems, which appears as a key recorder of the interaction of the main factors introduced above, and in particular records the interplay between high-frequency base-level and paleo-oceanographic fluctuations. Moreover, the presence of thick and expanded successions (clinothems developed up to hundreds of meters in thickness during intervals of only few thousands of years), allows to depict short-lived base-level and climatic variations at sub-Milankovitch time scales. Dramatic fluctuations in sediment supply coupled with high-frequency
base-level variations played a crucial role in determining stratal geometry of a river-dominated system by controlling sediment transport across the shelf, as well as phases of bypass off the shelf-edge and the timing of the storage of coarse-grained sediment in the basin (chapter 4). A short-lived increase in sediment flux during a still-stand that punctuated the post-glacial sea level rise accompanied by substantial lateral advection of sediment, played a fundamental role in determining an asymmetrical compound delta on the mid-shelf during the Younger Dryas, mimicking (in internal stratal geometry and spatial distribution) the modern Po River delta and its genetically-linked subaqueous prodelta system (chapter 5). Advection played also a crucial role on sediment routing along the outer shelf and slope of the southern Adriatic basin, promoting the development of shallow-water contourite deposits intermingled with prograding clinoforms (chapter 6).
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