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Post-glacial persistence of turbiditic activity within the Rhône deep-sea turbidite system (Gulf of Lions, Western Mediterranean): Linking the outer shelf and the basin sedimentary records

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ABSTRACT

Emplacement of post-glacial turbidites is commonly controlled by rapid changes in sea level or by seismicity. On the continental rise of the Gulf of Lions (Western Mediterranean), an aseismic area, we identified turbiditic beds deposited during the rising stage and highstand of sea level. Swath bathymetry, sediment cores, *in situ* Cone Penetrating Tests (CPTU), heavy mineral associations and radiocarbon dating determined the source, composition, distribution and age of the turbiditic beds. Turbidites are composed of homogeneous to positively graded silts to medium sand with quartz (up to 90%), shell debris and shelfal benthic faunas. Their distribution on the sea floor is very patchy and controlled by abundant inherited erosional bedforms. Their source is found in relict regressive sands at the outershelf. Their deposition occurred just after the onset of the post-glacial sea level rise and the concomitant sediment starvation of the Rhône deep sea turbiditic system until recently. Whilst canyons are fed with sand by strong seasonal hydro-sedimentary dynamics on the outershelf, the emplacement of post-glacial turbidites is not controlled by sea level changes but probably by the periodic flushing of the canyons. Our study revealed that this low energy aseismic margin undergoes significant transport of sand, down to the base of slope, during the sea-level rise and the Holocene highstand.
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1. Introduction

Sand exportation from the continents down to deep sea areas mainly occurs via canyons and turbiditic channels. This dominantly occurs during sea level lowstand stages, when river mouths used to be connected to canyon heads.

Other contexts and periods (cf. review of mass-transport deposits from Maslin et al., 2004) have been reported, notably in open environments, such as in abyssal plains, during times of sea-level change, e.g. the Madeira, Agadir and Seine abyssal plains (Weaver and Rothwell, 1987; Wynn et al., 2002b), and during the Holocene, e.g. in the Balearic Abyssal Plain (Rothwell et al., 2006), in the Horseshoe Abyssal Plain (Lebreiro et al., 1997) and the Balearic Abyssal Plain (Zuniga et al., 2007). This also occurs along continental slopes such as the Cascadia margin (Goldfinger et al., 2003) or the Markan continental slope (Prins et al., 2000b). In this context sand is transported by unconfined turbiditic flows that may initiate from the slope and which are possibly triggered by earthquakes (Lebreiro et al., 1997; Rothwell et al., 1998; Hieke and Werner, 2000; Goldfinger et al., 2003), and use adjacent canyons as a pathway. Examples of deep-sea high-stand coarse turbiditic deposits, despite no direct canyon–river connection, have been described in the Cap Timiris canyon (Wien et al., 2006), in the Indus canyon (Prins et al., 2000a), in the Bengal «Swatch of No Ground» canyon (Weber et al., 1997), on the Amazon fan (Flood and Piper, 1997), on the Celtic fan (Zaragosi et al., 2000), on the Armorican turbiditic system (Zaragosi et al., 2001), on the Hueneme and Dume fans (Normark et al., 1998). These deposits are interpreted as the product of reworked outer shelf or canyon head sands (Flood and Piper, 1997; Zaragosi et al., 2000, 2001), or of across-shelf sediment transport (Weber et al., 2003). They contrast with the classical concept of eustatic control of submarine fan development (Shanmugam et al., 1985), where the present flooding of the shelf would prevent the transfer of coarse sediment into the deep basin.

On the continental rise of the Gulf of Lions (western Mediterranean), at the foot of the canyon drainage network, in the vicinity of the Petit-Rhône deep-sea turbiditic system, (Fig. 1), several sediment cores sampled up to 30 silty and/or sandy layers, intercalated in hemipelagic deposits from the last glacial termination and from the Holocene. These layers were deposited during and after the last major sea-level rise. They overlay turbiditic spillover deposits in the distal part of the turbiditic channel of the Rhône deep-sea turbiditic system (Bonnel et al., 2005).

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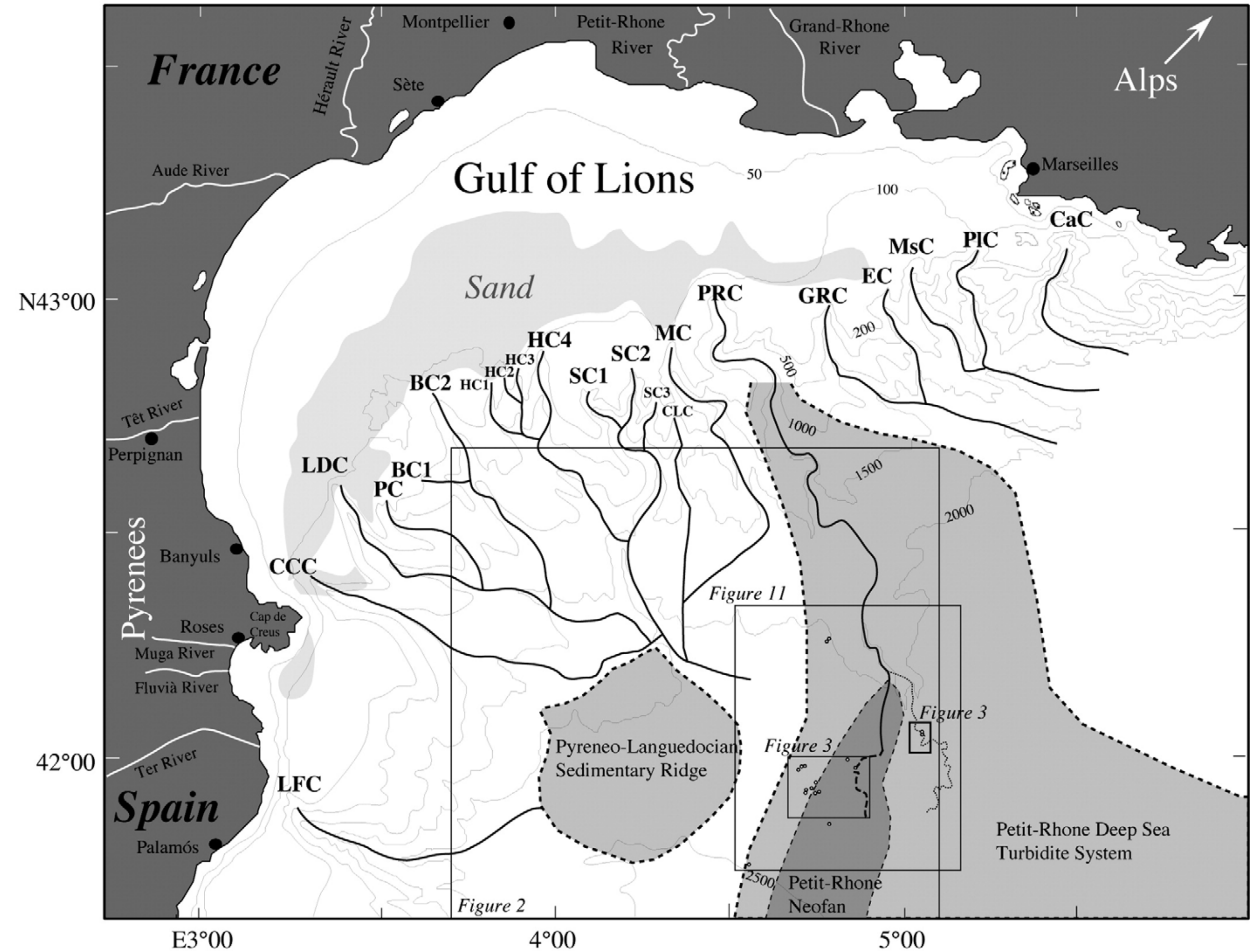

Fig. 1. Localisation and morpho-bathymetric map of the study area. LFC, La Fonera Canyon; CCC, Cap de Creus Canyon; LDC, Lacaze-Duthiers Canyon; PC, Pruvot Canyon; BC, Bourcart Canyon; HC, Hérault Canyon; SC, Sète Canyon; MC, Marti Canyon; PRC, Petit-Rhône Canyon; GRC, Grand-Rhône Canyon; EC, Estaque Canyon; MsC, Marseille Canyon; Plc, Planier Canyon; CaC, Cassidaigne Canyon. Numbering of sub-canyon is after Baztan et al. (2005). Distribution of sand on the shelf (very light grey) is after Aloïsi (1986) and Got and Aloïsi (1990). Position of the Petit-Rhône turbiditic system, Pyreneo-Languedocian Ridge (light grey) and Petit-Rhône neofan (grey) is after Droz et al. (2006). Localisation of Figs. 2, 3 and 11 is indicated. Open circles indicate sediment cores.

This paper deals with sedimentological evidences of post-glacial coarse sediment deposition and erosion at the base of slope in the Gulf of Lions during the last interglacial stage. We will focus on the distribution, composition, chronology and possible sources of these silt–sand layers. We will discuss the control of sea-level climate, physiography, and hydrography on the deposition of these layers.

2. Morpho-sedimentary setting

In the Gulf of Lions (western Mediterranean) the build up of the margin was strongly controlled by Quaternary glacial-interglacial sea-level variations and by significant subsidence at the shelf edge that led to the deposition and preservation of various types of sedimentary bodies and to numerous canyons dissecting the continental slope (Berné et al., 2001; Baztan et al., 2005) (Fig. 1).

The Rhône River is the main provider of sediments for the area and depot centres have moved across the margin concurrent to sea level changes. During the last sea-level rise and high stand sediments were stored in submarine prodeltas, still visible on the present seabed morphology, (Labaune et al., 2005; Berné et al., 2007) and partly redistributed to the west along the inner shelf through anti-clockwise circulation induced by general circulation and wind (Millot, 1991). During the last sea-level regression and low stand, sediments accumulated on the outer shelf, on the upper slope and the deep sea areas. On the outer shelf, between −110 m and −90 m, are deposits of littoral prograding shoreface sands (Bourcart, 1945; Monaco, 1971; Aloïsi, 1986; Berné et al., 1998, 1999; Jouet et al., 2006) of median grain size ranging from 200 µm to 500 µm) and composed of 25–50% of bioclastic carbonates. On the upper slope are deposits of fine-grained prodeltaic mud presently preserved on canyon interfluves (Rabineau et al., 1998; Tesson et al., 1990). During the low stand period, the base of slope collected most of the fluvial sediment which accumulated at the Pyreneo-Languedocian Ridge and the Rhône turbidite system (Fig. 1) (Méar, 1984; Droz and Bellaiche, 1985; dos Reis et al., 2004; Bonnel et al., 2005; Jallet and Giresse, 2005; Droz et al., 2006). The Pyreneo-Languedocian Ridge and the Petit-Rhône deep-sea turbiditic system have been growing concomitantly since the Middle Pleistocene (Droz and Bellaiche, 1985; dos Reis et al., 2004; Droz et al., 2006). On the western flank of the Rhône turbiditic system lies a Mass Transport Deposit, of probably the last glacial maximum, (Droz and Bellaiche, 1985; Méar and Gensous, 1993; Gaullier et al., 1998). Above, lie deposits from the last avulsion of the Rhône turbiditic channel called “neofan” (Droz and Bellaiche, 1985; Torres et al., 1997; Bonnel et al., 2005). Since 18.4 cal. ka BP spillovers of river-fed turbiditic flow

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