Different frequencies and triggers of canyon filling and flushing events in Nazaré Canyon, offshore Portugal

Joshua R. Allin[1,2], James E. Hunt[1], Peter J. Talling[1], Michael A. Clare[1], Ed Pope[1,2], Douglas G. Masson[1]

[1]National Oceanography Centre Southampton, European Way, Southampton, SO14 3ZH, UK

[2]University of Southampton, Waterfront Campus, European Way, SO14 3ZH, UK

Corresponding author: J. R. Allin (jra1g13@soton.ac.uk)

Abstract

Submarine canyons are one of the most important pathways for sediment transport into ocean basins. For this reason, understanding canyon architecture and sedimentary processes has importance for sediment budgets, carbon cycling, and geohazard assessment. Despite increasing knowledge of turbidity current triggers, the down-canyon variability in turbidity current frequency within most canyon systems is not well constrained. New AMS radiocarbon chronologies from canyon sediment cores illustrate significant variability in turbidity current frequency within Nazaré Canyon through time. Generalised linear models and Cox proportional hazards models indicate a strong influence of global sea level on the frequency of turbidity currents within the canyon. Radiocarbon chronologies from basin sediment cores indicate that larger, canyon-flushing turbidity currents reaching the Iberian Abyssal Plain have a significantly longer average recurrence interval than turbidity currents that fill the canyon. The recurrence intervals of these larger turbidity currents also appear to be unaffected by long-term changes in global sea level. This indicates that the factors triggering, and thus controlling, the frequency of canyon-flushing and canyon-filling events are very different. Canyon-filling appears to be predominantly controlled by sediment instability during sea level lowstand and by storm and nepheloid transport during the present day highstand. Canyon-flushing exhibits time-independent behaviour. This indicates that a temporally random process, or summation of non-random processes that cannot be discerned from a random signal, are triggering canyon flushing events.

1 Introduction

Understanding variability in turbidity current frequency and magnitude is important for several reasons. First, turbidity currents are one of the most voluminous sediment transport mechanisms, and they create some of the largest sediment accumulations on our planet (Ingersoll et al., 2003). Second, understanding the frequency and scale of large turbidity currents informs risk assessment for undersea installations that are at risk of damage by turbidity currents, such as oil and gas infrastructure, pipelines, and telecommunications cables (Bruschi et al., 2006; Carter et al., 2012; Carter et al., 2014).

Submarine canyon systems are recessed topographic features on continental margin slopes that act as conduits for sediment transport into the deep sea (Stow et al., 1985; Normark and Piper, 1991; van Weering et al., 2002). Turbidity currents are one of the main transport processes within submarine canyon systems, and can be triggered by a wide variety of mechanisms. Potential triggers include storm activity, tidal resuspension, sediment failures (triggered in some cases by earthquakes), and river discharges (Marshall, 1978; Masson et al., 2006; Piper and Normark, 2009; Masson et al., 2011a; Talling et al., 2012; Talling, 2014).

Turbidity currents in submarine canyons are proposed to be one of two broad end-member types: filling and flushing (Parker, 1982; Piper and Savoye, 1993; Canals et al., 2006; Piper and Normark, 2009; Talling et al., 2012). Canyon-filling turbidity currents are hypothesised to slowly deposit sediment within canyons over hundreds or even thousands of years (Paull et al., 2005; Canals et al., 2006; Arzola et al., 2008; Puig, et al., 2014). Canyon-filling turbidity currents are considered to be the result of localised sediment failures, hyperpycnal flows, or storm resuspension (Marshall, 1978; Arzola *et* al., 2008; Khripounoff et al., 2009; Masson et al., 2011a; Talling et al., 2013; Talling, 2014).

Canyon-flushing turbidity currents are erosive flows that remobilise and transport large volumes of this canyon-filling sediment out onto canyon-mouth fans or distal basin floors (Parker, 1982; Piper and Savoye, 1993; Xu et al., 2004; Paull et al., 2005; Piper and Normark, 2009; Kriphounoff et al., 2012; Talling et al., 2012; Puig et al., 2014). Canyon-flushing turbidity currents have yet to be directly monitored, and are suggested to operate on much longer timescales than those that fill the canyon (Piper and Savoye, 1993; de Stigter et al., 2007; Arzola et al., 2008; Talling et al., 2013). The causes of canyon-flushing events are not clear, although they likely result from large sediment failures (Normark and Piper, 1991; Masson et al., 2006; Goldfinger et al., 2007; Piper and Normark, 2009; Hunt et al., 2013a; Talling, 2014).

1.1 Observations of canyon-filling

Previous work on the Nazaré Canyon has focussed on the factors controlling sedimentation within the upper and middle reaches of the canyon (van Weering et al., 2002; de Stitger et al., 2007; Oliveira et al., 2007; Arzola et al., 2008; Martin et al. 2011). As many as four turbidity currents have been observed in the upper canyon per year (de Stigter et al., 2007; Martin et al., 2011; Masson et al., 2011a). These frequent turbidity currents are typically the result of winter storms that re-suspend canyon sediments, but may also be the result of small intra-canyon failures (de Stigter et al., 2007; Martin et al., 2011; Masson et al., 2011a). Direct monitoring from below 4,000 m water depth indicates that turbidity currents annually reach the lower canyon. However, these flows are generally dilute, restricted to the incised thalweg, and are not considered to be erosive (de Stigter et al., 2007).

Canyon terraces at 3500 m water depth and 40 m elevation above the thalweg) record multiple thicker (>25cm) turbidites deposited over the last 1,000 years. Turbidites recorded on terraces are interpreted as the result of large flushing events by Arzola et al. (2008) and Masson et al. (2011a). Given the presence of erosive scours within the lower reaches of the canyon, it is possible that these frequent turbidity currents are erosive and flush sediment onto the canyon-mouth fan (Arzola et al., 2008). To date, no deep water (>5000 m) core descriptions have been published from Nazaré Canyon, and so our understanding of the recurrence rates of both canyon-filling and canyon-flushing turbidity currents is limited to the upper and middle canyon, above 4,000 m water depth.

1.2 Aims of the study

Recent statistical analyses of long-term (>0.15 Ma) records indicates that large volume turbidites in distal basin plains can have a temporally random distribution, and are not strongly influenced by non-random glacio-eustatic sea level variability (Hunt et al., 2013a; Clare et al., 2014). Other studies suggest that sea level is a dominant control; however, they do not consider a sufficient number of events for statistical analysis (Maslin et al., 2004; Owen et al., 2007; Lee, 2009; Smith et al., 2013). It has also been proposed that climate-driven sea level change can affect the frequency of canyon-filling turbidity currents by influencing slope stability (Vail et al., 1977; Shanmugam and Moiola, 1982; Lebreiro et al., 1997; Piper and Normark, 2001; Lebreiro et al., 2009; Brothers et al., 2013). Therefore, we first aim to determine whether the frequency of canyon-filling shows a significant correlation with glacio-eustatic sea level change.

This present study is novel because few papers provide a complete overview of the frequency of turbidity currents from the upper to lower reaches of submarine canyons and out into deep water basins. No studies to date have statistically assessed both a basin and canyon record from a single system to test for any overarching control on the recurrences of canyon-filling and canyon-flushing. Therefore, also aim to assess whether or not canyon-flushing events have similar recurrence intervals to canyon filling events, and whether their recurrence intervals have similar statistical distributions. This will help determine whether they are likely to have similar (or different) triggers (Urlaub et al., 2013; Clare et al., 2014; 2015).

2 Regional setting

The study area is located on the Western Iberian Margin between 36^{o} and 43^{o} N (Fig. 1). The shelf and continental slope is incised by several large submarine canyons, including the Setúbal-Lisbon, Cascais, Sao Vincente and Nazaré canyons. These canyons feed into 3 deep sedimentary basins; the Iberian, Tagus and Horseshoe Abyssal Plains (Fig. 1).

2.1 Nazaré Canyon

Nazaré Canyon occurs in the central-west Iberian margin, and extends from ~1 km offshore from the coastline and into the Iberian Abyssal Plain (Fig. 1). The canyon incision into the continental shelf and slope coincides with the presence of the Nazaré Fault, which runs ENE-WSW and extends across the margin. The Nazaré Canyon system is not directly fed by major rivers, as is the case with the Setúbal and Cascais Canyons to the south. Instead, Nazaré Canyon is fed largely by littoral drift sediment from smaller river systems to the north, and nepheloid transport of material to deeper sections of the canyon (van Weering et al., 2002; Oliveira et al., 2007; de Stigter et al., 2007).

The canyon itself can be divided into three sections; the upper section that extends from the canyon head to 2,000 m water depth; the middle section that spans between 2,000 and 4,000 m; and the lower section that lies below 4,000 m (Figs. 2 and 3) (Vanney and Mougenot, 1990; van Weering et al., 2002). The upper and middle sections have a steep v-shaped profile and are incised deeply into the continental shelf (200 - 2,500 m water depth), with a channel thalweg that is <100 m wide. Below 4,000 m water depth the canyon broadens to a width of 8 – 10 km and is markedly less incised into the substrate. Below 4,500 m water depth the canyon is less incised, and large levee structures have developed to the north and south of the canyon axis. These levees are between 100 and 200 m in height and taper distally into the Iberian Abyssal Plain, where they terminate at 5,300 m water depth (Arzola et al., 2008; Lastras et al., 2009).

2.2 Iberian Abyssal Plain

The Iberian Abyssal Plain is located 200 km off the western coast of Portugal between 40^{o} N and 43^{o} N and extends approximately 700 km to the northwest. The basin plain has an average water depth of ~5,300 m but can be as deep as 5,400 m. The basin is bounded by the Galicia Bank to the northeast, the Estremadura Spur to the south, and by a series of seamounts along its western margin. The total area of the basin covers approximately 107,000 km^2. Previous work from ODP leg 149 has detailed the long-term basin infill record extending back to the Lower Cretaceous (140 Ma) (Milkert et al., 1996a; 1996b). This work demonstrated an onset of, terrestrial-derived turbidite deposition in the Iberian Abyssal Plain between 2.2 and 2.6 Ma, which continued into the late Pleistocene. Despite the extensive record from ODP drilling legs in the Iberian Abyssal Plain, very little of the recent (<100 ka) sedimentary architecture has been evaluated in detail. As a result, little is currently known about the frequency of large volume turbidity currents in the basin through the late Pleistocene.

3 Materials and methods

3.1 Piston coring

The cores used in this study have been collected by two different scientific campaigns. Piston cores JC27-51, JC27-47 and JC27-46 come from the abyssal plain and levee sections, and were recovered during cruise JC027. Core JC27-51 was retrieved from the Iberian Abyssal Plain, and cores JC27-46 and JC27-47 were retrieved from the northern external and internal levees respectively (Fig. 3). The piston cores from the middle section of Nazaré Canyon were retrieved during cruise CD157. Cores were sited using bathymetric and geophysical data. Cores D15738 and D15739 were both collected from terraces 40 - 60 m above the canyon axis and have existing radiocarbon dates (Figs. 2 and 3).

There are no sediment cores from the upper canyon above 2,000 m water depth from which AMS radiocarbon ages could be determined. The coarse-grained nature of the sediment in the upper canyon, the rapid rate of accumulation, and the tidally-driven re-working of material prevent any long-term record of turbidity currents being obtained through piston or gravity coring techniques (de Stigter et al., 2007).

3.2 Identification of turbidites and hemipelagite

In order to develop age models and determine recurrence intervals for turbidity currents, differentiating their deposits from background hemipelagic sediment is vital. Hemipelagic sediments typically contain randomly dispersed foraminifera giving a pitted surface texture. They also lack primary sedimentary structures and are often bioturbated (Stow and Piper, 1984). In contrast, turbidity current deposits are often well sorted, have normal grading, and have observable internal structure. The fine-grained mud cap of turbidity currents is commonly homogenous and often devoid of foraminiferal material, while the basal contact of turbidites is often sharp and sometimes erosional (Bouma, 1962: Stow and Piper, 1984).

Cores were logged for sedimentary features such as grain size, sedimentary structures and colour. The cores were logged using a Geotek™ multi-sensor core logger (MSCL) for petrophysical data, particularly optical lightness and spectral reflectance. ITRAX-XRF geochemical data were obtained for JC27-51 to aid in the construction of the age-model using hemipelagic sediment. It has been shown that hemipelagic sediment in basin settings has a chemistry that is distinct from terrestrially-derived siliciclastic sediment, primarily due to a high proportion of detrital carbonate (Croudace et al., 2006). Where ITRAX-XRF data were not available, optical lightness (L*) was used to discriminate between Turbidite and hemipelagic deposits. High amounts of foraminiferal carbonate in basin sediments result in a lighter colour than terrestrially derived mass-flow deposits (Balsam et al., 1999).

3.3 Age model development

Age control for this study is provided by AMS radiocarbon dating. AMS radiocarbon dates were obtained for cores JC27-51, JC27-47 and JC27-46 in order to develop age models and calculate turbidity current recurrence intervals. Nine AMS radiocarbon dates were previously determined for terrace cores D15738 and D15739 (Masson et al., 2010) (Table 1). Samples were taken from hemipelagic mud units, with care being taken to avoid sampling any turbidite mud or coarse turbidite bases. In cores JC27-51, JC27-47 and JC27-46, 2 - 10 cm^3 of sediment was sampled to pick the 8 - 10 mg of foraminifera required for accurate AMS 14C dates. No one foraminiferal species was abundant enough to collect monospecific samples in all cases, so most samples consist of mixed species assemblages. The dominant species were *Orbulina universa, Globigerina bulloides, Neogloboquadrina pachyderma, Globorotalia truncatulinoides, Globigerinoides ruber* and *Globorotalia hirsute*. Where monospecific samples were possible, *Orbulina universa* was selected. The conventional radiocarbon ages returned from analysis were converted to calibrated ages (Cal years BP) using the MARINE13 database (Reimer et al., 2013) (Table 1).

In order to account for local reservoir offsets, an average of five reservoir correction (ΔR) values was used from nearby locations. These ΔR values are from samples collected along the Iberian shelf; the main pathway for ocean currents and sediment transport to the head of Nazaré Canyon (Monges Soares, 1993; van Weering et al., 2002; Oliveira et al., 2007; de Stigter et al., 2007). This yields a ΔR correction value of +267 years, which is consistent with reconstructed past reservoir offsets along the Iberian Margin (Bronk Ramsey et al., 2012).

Using the radiocarbon ages, and the thicknesses of hemipelagic sediment between them, sedimentation rates were calculated. The thicknesses of hemipelagic sediment were then divided by the sedimentation rates to convert them into time intervals (Wynn et al., 2002; Grácia et al., 2010; Clare et al., 2014). From these time intervals the ages of individual turbidites can be estimated (Table

2; Suppl. info.). This method relies on the assumption that there is minimal fluctuation in the rate of hemipelagic sediment accumulation through time (Lebreiro et al., 2009; Grácia *et al*, 2010; Clare et al., 2015). It also relies on the assumption that subsequent turbidity currents are not significantly erosive (Weaver and Thomson, 1993; Thomson and Weaver, 1994; Weaver, 1994; Wynn et al., 2002; Gutiérrez-Pastor et al., 2009; Grácia et al., 2010).

3.4 Calculation of turbidite recurrence intervals and frequency

Using the age model to estimate the emplacement age of each turbidite allows us to calculate individual recurrence intervals. Here we define the recurrence interval of a turbidite as the length of time since the turbidite that preceded it (Clare et al., 2014; 2015; Pope et al., 2015). Where hemipelagic age models cannot be constructed, we calculate recurrence interval by diving the length of time by the number of turbidites to get an 'average recurrence interval'. The determination of recurrence intervals requires that individual turbidites be distinguished from multiple upward-fining units deposited during the same turbidity current. These multiples of upward-fining sediment can result from differential sorting due to changing bed-shear stresses, or from multi-staged failures. (Piper and Bowen, 1978; Stow and Shanmugam, 1980; Hunt et al., 2013b). Interpreting multiple upward-fining units within a single turbidite as multiple individual turbidites has the potential to bias any analysis by the incorrect counting of turbidites; and hence, incorrectly estimating their recurrence intervals (Lebreiro et al., 2009).

3.5 Statistical analysis of turbidite recurrence and frequency

Statistical analysis can be a powerful tool for the analysis of time series data, such as turbidite recurrence (e.g. Hunt et al., 2014; Moernaut et al., 2015; Ratzov et al., 2015); however, it is important to understand how recurrence is measured before specific tests are selected. Recurrence is here inferred from intervals of hemipelagic fallout between turbidity currents, and the average accumulation rate of hemipelagic mud between dated horizons. This method is most appropriate in distal basin settings, such as basin core JC27-51 where past work has shown there is little or no erosion by successive turbidity current, and the hemipelagic mud is not removed (Weaver and Thomson, 1993; Thomson and Weaver, 1994; Weaver, 1994; Wynn et al., 2002; Gutiérrez-Pastor et al., 2009; Grácia et al., 2010; Clare et al., 2014). In more proximal, slope and confined settings, the effects of erosion may be greater, but it is difficult to discern this from core samples due to their relatively narrow diameter (typically 10 cm). We therefore bin the data at the more proximal levee core site JC27-46, by counting the number of turbidites within prescribed time intervals to account for this uncertainty in the precise measurement of individual recurrence intervals. As bin dimensions can potentially affect the statistical outcome (Urlaub et al., 2013; Pope et al., 2015) we consider three different bin widths (250, 500 and 1000 years) in our analysis. We investigate the influence of sea level on binned turbidite recurrence at levee site JC27-46, where sufficient turbidites were sampled (N=201) to permit regression and survival analysis. In addition to different data bin sizes, we also use three different sea level curves as our explanatory variable. The sea level reconstructions of Lambeck et al. (2014), Rohling et al. (2010), and Peltier and Fairbanks (2006) are widely used and have suitable resolution over the late Pleistocene. Less sophisticated frequency analysis is used for individual turbidite recurrence intervals at basin site JC27-51, as fewer turbidites were sampled (N=26) thus limiting the power of statistical tests (Pope et al., 2015).

3.5.1 Assessing the significance of sea level on turbidite recurrence at JC27-46

We use two different statistical analyses to investigate the significance of sea level in relation to turbidite recurrence at JC27-46. The first is a parametric Generalised Linear Model, which tests for the significance of sea level as an explanatory variable on the recurrence of turbidites (McCullagh and Nelder, 1989; Clare et al., 2016). As it is a parametric test, the Generalised Linear Model requires *a priori* definition of the distribution form of the relationship between the two variables. We apply variants of the model using Gamma, Gaussian and Poisson distribution forms, and assess the most appropriate model using quantile-quantile (Q-Q) plots. While Q-Q plots do not provide a quantitative measure of the goodness-of-fit, they can be qualitatively interpreted to understand how much a given data set deviates from the specified distribution (Salkind and Rasmussen, 2007).

General rules of thumb exist for determining the minimum sample size for regression analysis. Tabachnick and Fidell (2007) indicate that testing for the effect of one individual variable will require N=106 events. Green (1991) performed a more detailed analysis, which incorporated an assessment of statistical power and effect size, suggesting that at least N=23 is required to detect large effects and N=53 for detecting medium effects of one explanatory variable. We also apply a non-parametric Cox Proportional Hazards (PH) Model (Cox, 1972) as a comparative test because it requires no *a priori* specification of frequency distribution form. The Cox PH model is typically used to determine a hazard rate in medical studies (e.g. rate of patient fatality) but has also been applied to turbidite frequency analysis (Hunt et al., 2014; Clare et al., 2016). The hazard rate is the ratio between the change in the explanatory variable (e.g. sea level) and the response variable – in this case turbidite recurrence. Previous work has shown that the Cox PH model requires at least a minimum sample size of N=20 (Vittinghoff and McCuloch, 2007). The Cox PH model performs survival analysis and three separate tests (likelihood, Wald and log-rank), for which a p-value is derived. For both Generalised linear and Cox PH models, where the resultant p-value is small (<0.05) sea level is found to be a significant variable to explain turbidite recurrence. Where the p-value is large ($p>0.05$), sea level cannot be implicated as a statistically significant control on turbidite recurrence. Levee core site JC27-46 features N=201 turbidites, which is well above the minimum sample size threshold for both regression and survival analysis. Distal basin core JC27-51 only features N=26 turbidites, which is at the minimum sample size limit, and therefore would yield only low statistical power. Therefore, we simply analyse the frequency distribution form of turbidite recurrence at JC27-51 to provide insights into triggering and possible controls, instead of performing more sophisticated survival or regression analysis.

3.5.2 What is the frequency distribution form of turbidite recurrence at JC27-51?

A Poisson distribution implies time-independence and a lack of memory in a system (Parzen, 1962). Thus, recurrence intervals that fit a Poisson (exponential) distribution form can be viewed as occurring randomly in time, with no dependence on when the previous event occurred, or when the next will happen. This memoryless, time-independent behaviour is in contrast to non-random processes such as sea level change. Therefore, where recurrence intervals conform to a Poisson distribution, non-random processes cannot be directly attributed to a singular or dominant control on recurrence (Urlaub et al., 2013; Clare et al., 2014). In contrast, time-dependent distributions may indicate that a single process (e.g. Normal) or series of processes (e.g. log-normal) are exerting significant control on a system (van Rooij et al., 2013; Clare et al., 2016). Therefore, we aim to determine the frequency distribution form of turbidite recurrence intervals at JC27-51 to provide some insights into possible triggering and controlling mechanisms for canyon flushing events. We also aim to determine whether the frequency distribution form of turbidite recurrence at the basin plain is distinctly different to that observed at the more proximal levee location.

Here we employ parametric and non-parametric goodness-of-fit methods to test recurrence interval distribution. The Anderson-Darling test is a parametric test that tells us the probabilities that the data come from different populations with specific distributions, including Poisson, normal, log-normal and Weibull (Stephens, 1974). To further test the distribution of canyon flushing events, we use non-parametric (Mann-Whitney and Kolmogorov-Smirnov) tests to determine if the frequency distribution form of Iberian Basin turbidite recurrence is significantly different to that of other distal basin plains. For this, we use the published dataset of Clare et al. (2014; 2015), which demonstrates that multiple basin turbidite records conform to a Poisson distribution. The Mann-Whitney test is based on the null hypothesis that the datasets are sampled from populations with identical distributions (Lehmann and D’Abrera, 2006). The Kolmogorov-Smirnov test compares the cumulative distributions of two data sets and poses the null hypothesis that they were randomly sampled from populations with identical frequency distributions (Lehmann and D'Abrera, 2006). The result tells us the probability that the two cumulative frequency distributions would be as far apart as observed in our data. We use these three tests to provide confidence in our results.

4 Results

4.1 Core sedimentary characteristics

Core JC27-51 shows a depositional sequence comprised of pale grey mud units with abundant foraminifera interbedded with well sorted and normally-graded sedimentary units (Fig. 3). The graded units are often greater than 50 cm thick and many are mud-dominated (Fig. 4A). Some of these mud-dominated deposits exhibit thin silt or fine sand bases. Three deposits within the core have sandy bases, with one of these (core depth 320 - 390 cm) having a much thicker medium- to coarse-grained sand base with planar and ripple laminations (Fig. 4B). We interpret these graded units as turbidite deposits after Bouma (1962), and Stow and Piper (1984). A silt unit with interspersed mud clasts occurs at 395 - 420 cm depth in the core, and has a subtle reverse-to-normally graded sequence (Fig. 4C), which we interpret as a debrite, in the sense of Naylor (1980).

The pale sediment in core JC27-51 exhibits different colour, grain size, and lightness characteristics than turbidite deposits (Fig. 5A). Unlike the more homogenous olive-green turbidites, the pale grey sediment is noticeably rich in foraminiferal sand and exhibits bioturbation structures. This foraminiferal sand gives the sediment a pitted surface texture. From the photospectrometer data the pale grey sediment in JC27-51 exhibits an L* value typically within a range of 50 - 65, while turbidites are typically within 35 - 50. From the ITRAX XRF Calcium counts (kcps) data, the pale grey sediment typically exhibits values greater than 160 kcps (Fig. 5A). In contrast, turbidite deposits have values ranging from 40 to 120 kcps. These differences in the geochemical and geophysical properties are likely due to the high amounts of calcareous foraminiferal shells within the sediment. We interpret this pale grey sediment as background hemipelagic accumulation, after Stow and Piper (1984) and Hoogakker et al. (2004). Typically, the boundaries between the upper turbidite mud cap and the following hemipelagic sediment are gradational (Fig. 5A). This is due largely to bioturbation, but also likely due to the slow settling of turbidite mud suspension clouds and incorporation of carbonate material from background hemipelagic sediment. The boundaries between the bases of the turbidite deposits are flat and sharp (Fig. 5A).

Core JC27-46 consists largely of interbedded silt and fine sand units that are normally graded and typically have thin mud caps (Fig 4). These graded deposits can be broadly classified into two different types. Type 1 deposits are normally graded, with 1 - 4 cm thick silt or fine sand bases, and thicker 10 - 20 cm upper mud units. Type 2 deposits are thinly-bedded and have 0.5 - 4 cm thick, faintly gradational fine- to medium-grained sand bases. The thin sand bases typically grade sharply into a thin (0.5 - 3 cm) fine-grained mud (Fig. 4D and E). Type 1 deposits are dominant in the upper 6 m of the core, while type 2 are dominant in the lower 4 m. We interpret these deposits to be the result of turbidity currents in the sense of Stow and Shanmugam (1980). Within the core there are lighter, pale grey units that are rich in foraminiferal sand, have a pitted surface, and show evidence of bioturbation. Based on the similarity with hemipelagic deposits from core JC27-51 we also interpret these to have a hemipelagic origin.

Core JC27-47 is dominated by thicker sand-rich, normally-graded deposits with medium- to coarse-grained sand bases (Fig. 4F and G). Sand units vary between 0.5 and 20 cm thick and grade sharply into fine-grained mud. Several of these deposits have clear erosional bases, with several units presenting a chaotic texture and uneven or folded bedding (Fig. 4G). We interpret deposits with normal grading and erosional bases to be turbidites (Fig. 4F). Deposits with folded bedding or chaotic texture are interpreted to be the result of small-scale slumping or debris flows after Shanmugam et al. (1995). The upper 60 cm of the core contains of the same pale grey, bioturbated and foraminiferal-rich sediment present in cores JC27-46 and JC27-51. As with the two previous cores, we interpret this to be hemipelagic sediment. Hemipelagic sediment is not present below 60 cm depth in the core, possibly due to erosion from subsequent turbidites over-spilling the internal canyon levee (Fig. 3).

Middle canyon cores are located on terraces at 40 - 60 m elevation above the canyon thalweg (Fig. 3). Cores D15738 and D15739 are composed largely of clay with interspersed faintly-graded fine sand or silt beds (Fig. 3). These are interpreted by Arzola et al*.* (2008) as turbidite deposits resulting from turbidity currents that overspill onto canyon terraces. Basal contacts between the sand and the underlying mud are typically erosional, and previous work indicates there is no discernible hemipelagic material between turbidite units due to the high level of shelf-derived terrigenous material transported in nepheloid layers to deeper in the canyon (de Stigter et al., 2007; Arzola et al., 2008).

4.2 Age model and sedimentation rate

Seven AMS radiocarbon dates were collected from JC27-51, five from JC27-46 and five from JC27-47 (Fig. 3). Using these dates and the known thickness of hemipelagic sedimentation, age models place the base of core JC27-51 at ca 83,000 Cal years BP and the base of core JC27-46 at ca 33,500 Cal years BP (Fig. 6). Linear regressions on the age models yielded R^2 values of 0.9934 and 0.9975 in cores JC27-51 and JC27-46 respectively (Fig. 6). These values give us a high level of certainty in a stable hemipelagic sedimentation rate over long periods at both core sites (Swan and Sandilands, 1995). Average hemipelagic sedimentation rates for cores JC27-51 and JC27-46 are 2.3 and 7.8 cm/ka respectively (Fig. 6).

Individual turbidite ages derived from the two age models can be seen in Table 2, and in Supplementary Table 1. Due to the presence of multiple erosive sandy turbidites in core JC27-47, hemipelagic material is only discernible in the upper 60 cm (Fig. 3). Below 60 cm core depth, hemipelagic sediment is sparse and indistinguishable from fine-grained turbidite deposits. Similarly in cores D15738 and D15739, hemipelagic deposits contain a high percentage of terrigenous mud with sparse foraminifera, and are therefore indistinguishable from turbidite muds (Arzola et al., 2008; Masson et al., 2011a). AMS radiocarbon dates from cores JC27-47, D15738 and D15739 originate from foraminifera collected out of bulk sediment samples. Because of the uncertainty in correctly defining hemipelagic thicknesses, recurrence intervals for each deposit cannot be calculated using a hemipelagic age model. Instead, an average turbidity current recurrence interval is calculated by dividing the number of turbidites by the time between radiocarbon dates (Fig. 3).

4.3 Regular filling events in the middle canyon

Previously collected AMS radiocarbon ages from middle canyon terrace cores place the bases of cores D15738 and D15739 at ca 1,500 BP and ca 1,000 BP respectively. Core D15738 contains 25 identifiable sandy turbidite bases above the oldest AMS radiocarbon date, while core D15739 has 21 turbidites above the oldest radiocarbon date (Fig. 3). By dividing the time interval by the number of turbidites we are able to calculate an average recurrence for turbidites. This gives us give an average turbidity current recurrence of 43 years on terraces that are 40 - 60 m above the canyon floor.

4.4 Sea level control on canyon-filling in JC27-46

For visual comparison of sea level and turbidite frequency, we use the 500 year binned turbidite data from core JC27-46 (Fig. 7A). This 500 year interval is considered to be an appropriate resolution for the length of the record (Lebreiro et al., 2009). Turbidite frequency in core JC27-46 on the external levee crest is highly variable during the sea level lowstand from 33.5 to 23.5 ka; the 10,000 year period prior to North Atlantic warming and associated sea level rise (Clark et al., 2004; Anders and Carlson, 2012) (Fig. 7A). From the modelled ages for turbidites, an average of the estimated recurrence intervals during lowstand conditions is 68 years, lower than at any point during the 33,500 year record.

Turbidite frequency decreased to 0 – 3 events per 500 years during the period from 23 and 19.5 ka, the onset of sea level transgression (Fig. 7A). This initial decline of turbidite frequency starting at 23 ka occurred considerably before the rapid global eustatic sea level rise at 20 ka (Chappelle, 2002; Peltier and Fairbanks, 2006; Clark et al., 2009; Rohling et al., 2009; Lambeck et al., 2014). Following the onset of rapid sea level rise at 17 - 19 ka (Clark et al., 2004; 2009), there is an increase in frequency (4 turbidites/500 years) that lasts until 15.5 ka. Turbidite frequency decreased significantly after 15 ka towards the Younger Dryas. An average of recurrence intervals for the global sea level transgression is 300 years. Following the onset of the present-day sea level highstand at ~7 ka there is an average of recurrence intervals of 1625 years (Fig. 7A).

A similar pattern of increasing recurrence intervals through the period of deglaciation can be seen in JC27-47 (Fig. 7B). The recurrence interval values for turbidites are also broadly comparable to those recorded in JC27-46 on the external levee for the same time period. In JC27-47, recurrence intervals are approximately 120 years at the beginning of sea level transgression at 20 ka. During the end of the sea level transgression the average recurrence interval increases to 320 years, similar to the 300 year recurrence rate found in JC27-46 (Fig. 7A and B). As in core JC27-46, following the onset of the present-day highstand at 7 ka, recurrence intervals have increased to >1000 years (Fig. 7B).

The results of generalised linear models indicate that sea level is a significant explanatory variable ($p<0.05$; Supplementary table 2) in relation to the frequency of late Pleistocene canyon-filling turbidity currents at JC27-46. As observed visually, periods of lowered sea level correspond to more frequent turbidity currents. This correlation holds for individual and binned (250, 500 and 1000 year) recurrence intervals, against all of the sea level curves that are considered by this study (i.e. (Rohling et al., 2010; Peltier and Fairbanks, 2006; Lambeck et al., 2014; Suppl. Table 2). A Gaussian distribution appears to best parameterise the relationship. Cox Proportional Hazard Models also indicate a significant influence of sea level on the turbidite recurrence for all but two model runs; a significant relationship was not observed when the Lambeck et al. (2014) sea level curve is analysed against 250 year-bins and individual turbidite recurrence (Suppl. table 2).

4.5 Time-independent canyon-flushing in JC27-51

The most recent turbidite present in core JC27-51 from the Iberian Abyssal Plain dates to ca 4,850 Cal years BP, while the oldest dates to ca 82,000 Cal years BP (Table 2). Based on a hemipelagic sedimentation rate of 2.3 cm/ka (Fig. 6) a lack of hemipelagite between two turbidites indicates a short recurrence interval. Turbidity currents in the basin are assumed to be non-erosive, and so two or more turbidites with no intervening hemipelagite are assumed to be near-synchronous, and are assigned a recurrence interval of 0 years (Table 2). This gives a 2,880 year average of recurrence intervals for flushing turbidity currents that reach the Iberian Abyssal Plain.

We compare the recurrence intervals for JC27-51 with data from four other distal basin plains (Clare et al., 2014; 2015). When recurrence intervals are normalised to the mean value for each dataset, they all fit an approximately straight line fit on an exceedance plot (Fig. 8), which is indicative of an exponential distribution. Non-parametric Mann-Whitney and Kolmogorov-Smirnov tests demonstrate statistically that the distribution form of turbidite recurrence at JC27-51 shows no significant difference to the four other exponentially-distributed basin plain records (Supplementary Table 3). Given the relatively small sample size (i.e. N=26), it is possible that we do not have sufficient turbidites to make a fully conclusive statement, but we can state that it is not possible to differentiate the recurrence record from other time-independent records. The Mann-Whitney and Kolmogorov-Smirnov tests also allow us to demonstrate that the frequency distribution form of turbidite recurrence in the basin plain (JC27-51) is significantly different ($p<0.0001$) to the time-dependent recurrence at levee location JC27-46. Thus, recurrence intervals for canyon-flushing turbidites in the Iberian Abyssal Plain may be considered time-independent and occurring at a significantly different tempo and to that of canyon-filling flows.

5. Discussion

In this section we first discuss the triggers and controls on canyon-filling turbidity currents in Nazaré Canyon, and how external factors such as sea level affect recurrence rates. Second, we discuss the possible triggering mechanisms for large canyon-flushing turbidity currents. Finally, we evaluate the implications of these findings for geohazard assessment and in light of future climate change predictions.

5.1 Canyon-filling in the present day highstand

The process of canyon-filling occurs on a variety of different temporal and spatial scales within Nazaré Canyon. Previous monitoring in Nazaré Canyon has documented as many as four turbidity currents may occur in the upper canyon per year (de Stigter et al., 2007; Martin et al., 2011; Masson et al., 2011a). These small and frequent turbidity currents do not appear to deposit on canyon levees, or the abyssal plain, as there is a general absence of turbidites deposited more recently than 3 ka (Figs 3, and 7A and B). These sub-annual turbidity currents are confined primarily to the incised canyon thalweg and typically dissipate before reaching depths greater than 4,000 m water depth (Fig. 9A). One turbidity current recorded by de Stigter et al. (2007) below 4,000 m water depth also did not appear to deposit on canyon levees or on the abyssal plain. These flows may erode previously emplaced sediment, but are not considered to flush large volumes of material (Fig. 9A).

Terrace cores at 3,500 m water depth and 40 m elevation above the thalweg reveal a mean recurrence of 43 years for larger turbidity currents over the last 1,000 years. These distal turbidite deposits are interpreted as the result of large flushing turbidity currents by Arzola *et al.* (2008), although there is no evidence of post 1 ka deposits in the Iberian Abyssal Plain that would indicate canyon-flushing has occurred (Fig. 4, and 7A and B). Given the presence of erosive scours within the lower reaches of the canyon, it is possible that these frequent turbidity currents are erosive and do flush sediment from for the upper to the lower canyon, but are too small to overtop levees or reach the abyssal plain (Fig. 9A) (de Stigter et al., 2002; Arzola et al., 2008). However, no age-models exist for lower canyon cores making this difficult to support.

5.2 Canyon-filling during sea level lowstand

The role of sea level lowstand in exposing continental shelves and resulting in more terrigenous sediment delivery to slope and deep-sea fans is widely accepted; as is the role of canyons as conduits for this sediment delivery (Vail et al., 1977; Shanmugam and Moiola, 1982; Posamentier et al., 1991; Piper and Savoye, 1993; Lebreiro et al., 1997; Clark and Mix, 2000; Ducassou et al., 2009; Lebreiro et al., 2009; Covault and Graham, 2010). Our hemipelagic age model reveals that the frequency of canyon-filling turbidity currents was highest during sea level lowstand (Fig. 7A and B). Lowstand-dominated canyons typically occur when the canyon does not incise the entire continental shelf (Covault and Graham, 2010). During highstand conditions such canyons are not in direct contact with the shoreline or fluvial sources, limiting sediment delivery. During sea level lowstand, sequence stratigraphic models predict direct connection between rivers and canyon heads. This results in a higher frequency of turbidity currents due to increased sediment supply and associated instability (Shanmugam and Moiola, 1982; Stow et al., 1984; Posamentier et al., 1991; Piper and Savoye, 1993; Lebreiro et al., 2009; Covault and Graham, 2010).

During sea level lowstand the larger area of exposed continental shelf likely resulted in greater erosion and littoral sediment delivery to the Nazaré Canyon head (Fig. 9B) (Posamentier et al., 1991; Sommerfield and Lee, 2004; Durán et al., 2013). Despite the lack of a direct fluvial sediment supply into Nazaré Canyon, there are a number of small river systems to the north which feed onto the continental shelf (Fig. 9B). Increased sediment supply to the shelf from these river systems during lowstand conditions may also have led to increased littoral sediment transport, and contributed to the higher frequency of turbidity currents being generated (Fig. 9B). It is also possible that these river systems were routed across the shelf and into the canyon head during sea level lowstand, although there is no geomorphic evidence available to support this.

Lebreiro et al. (2009) have suggested that slope instability in the Setúbal/Lisbon and Cascais Canyons (Fig. 1) is highest during sea level transgression. This implies that large turbidity currents that fill the Setúbal/Lisbon and Cascais Canyons switch-off at 15.5 ka. Work by Masson et al. (2011b) instead demonstrates that the Setúbal/Lisbon and Cascais Canyons experience an almost complete switch-off in large canyon-filling turbidity current activity at 6.6 ka. The switch-off of canyon-filling activity in the Setúbal/Lisbon and Cascais Canyons proposed by Lebreiro et al. (2009) at 15.5 ka could more easily be explained by the 200 m elevation of their core site. This elevation above the canyon would bias the record towards the largest turbidity currents (de Stigter et al., 2011). Our age models and statistical analyses from Nazaré Canyon also indicate that large canyon-filling turbidity currents switch-off at ~7 ka (Fig. 9B). This is suggestive of a margin-wide sea level control on large-scale canyon-filling in Portuguese Margin canyons.

5.2.1 Uncertainties of the statistical analyses

The results of the generalised linear models and Cox proportional hazards models indicate that sea level is well correlated with turbidite frequency recorded on the external canyon levee (Supplementary table 2). While sea level is generally accepted as having a dominant control on sedimentation in many other deep sea fan settings, our results should be treated with some caution. Our data set contains 201 turbidites, and while this is a sufficient number for regression-based analyses (Green, 1991; VanVoorhis and Morgan, 2007), it is smaller than many other datasets used in statistical analyses of recurrence (Clare et al., 2014; 2015). The height of the JC27-46 core site above the canyon floor may also bias the turbidite record towards those turbidity currents large enough to over-spill the outer levee. There may also be turbidite deposits that are not visually detectable because they are sub-millimetre in scale. Moreover, the sea level curves used in this study are all global (eustatic) reconstructions, and may not be suitable for local analysis where the effect of eustatic sea level changes may be outweighed or convoluted by localised isostatic and tectonic influences (Shanmugam and Moiola, 1982; Stow et al., 1984; Covault and Graham, 2010; Romans et al., 2015). Importantly, binning the data appears to have an effect on the significance of sea level. 500 year and 1000 year bin sizes appear to be less significant or not significant when tested against sea level (Supplementary Table 2). This has also been observed in other statistical analyses of turbidite recurrence (Urlaub et al., 2013; Pope et al., 2015)

5.3 Potential triggers of canyon-flushing

The frequency distribution of canyon flushing flows in Nazaré Canyon is indicative of temporally random, time-independent behaviour. A temporally random distribution can result from *a)* a single temporally random or pseudo-random process, *b)* numerous non-random processes affecting a single source overprinting and resulting in a random distribution, *c)* several different sources feeding into the same basin, or *d)* shredding of environmental (triggering) signal due to the long recurrence intervals (Urlaub et al., 2013; van Rooij et al., 2013; Clare et al., 2014; 2015; Pope et al., 2015; Romans et al., 2015). Here we will evaluate different processes that could trigger canyon-flushing events.

5.3.1 Are large storms a trigger?

It has been stated that storms have the potential to trigger turbidity currents that transport sand into deep water (TsuTsui et al., 1987; Shanmugam, 2008). Within Nazaré Canyon, storm-triggered turbidity currents play a significant role in small-scale canyon-filling during the present day (Fig. 9A) (van Weering et al., 2002; de Stigter et al., 2007; Martin et al., 2011; Masson et al., 2011a). One of the limiting factors in assessing the role of storms in canyon-flushing events is the length of the observational record. Meteorological data available for the last 63 years indicates that there have been over 150 storms (wave height >1.6 m) recorded since 1952 (Lozano et al., 2004; Almeida et al., 2011). Despite the number of storm events, we see no recent turbidite deposits in the levee or abyssal plain cores; however, the cores may not contain the most recent sediments due to sampling loss. If storms had been a trigger of canyon-flushing through the Holocene, we would expect to see turbidites throughout Holocene sedimentary sequences. It may be possible that some extremely rare storms are large enough to generate turbidity currents that are erosive and capable of flushing the canyon. The lack of turbidites observed in the Iberian Abyssal Plain during last 2,000 years, combined with the likely number of storm events, makes it improbable that storms are a trigger of canyon-flushing (Fig. 3 and table 2).

5.3.2 Is sea level control undetectable due to 'signal shredding'?

Certain sedimentary environments serve as ideal long-term archives of climatic or environmental signals. This is particularly true in the case of deep-water fan and continental slope settings. In these deep-water settings, one of the principal expressions of sea level variability is changes in down-system sediment transport (Stow and Piper, 1984; Posamentier et al., 1991; Covault and Graham, 2010; Covault et al., 2010).

Signal shredding can be defined here as the filtration of environmental signals through a system by a non-linear process; in this case sediment transport (Jerolmack and Paola, 2010; Romans et al., 2015).

Environmental signals like sea level may be considered a linear input into a system. However, the recording of this signal in sedimentary archives by sediment transport and deposition is typically non-linear. This is because sediment transport and deposition is highly variable and dependent on several initial and transport conditions. For turbidity currents these can include the triggering mechanism, sediment volume, run-out distance, and whether or not these currents bypass or deposit (Piper and Normark, 2009; Talling et al., 2013, Talling, 2014). These preconditions impart non-linearity that can ultimately filter or 'shred' the signal of environmental change. In the case of Nazaré canyon-filling turbidity currents, the sea level signal is preserved in the levee due to the number of events recorded, the proximity to source, and the regularity with which they were deposited. The height of the levee above the canyon also likely prevents smaller, more frequent events triggered by storm or wave activity from depositing at the core site. Storm and wave activity could be considered a separate environmental signal and may mask the longer-term sea level signal (Romans et al., 2015).

In contrast to this, canyon-flushing events are typically less frequent, often having recurrence intervals of >2000 years (Fig. 9 and Table 2). They involve the remobilisation of sediment stored within the canyon over many thousands of years (Paull et al., 2005; Talling et al., 2007; Piper and Normark, 2009; Talling, 2014). This implies that if canyon-flushing events are the result of climatic variability or change, their rarity (non-linearity) in the depositional record could have shredded any signal of this climatic control (Covault and Fildani, 2014; Clare et al., 2015; Romans et al., 2015). Such signal-shredding could explain the temporally random distribution of canyon-flushing events in the Iberian Abyssal Plain.

5.3.3 Are regional earthquakes a trigger?

Turbidite paleoseismology involves the use of marine or lacustrine turbidite recurrence rates as a proxy for earthquake recurrence. This method has been applied to numerous settings to develop estimates of long-term earthquake hazard rate (Adams, 1990; Monecke et al., 2004; Goldfinger et al., 2007; Polonia et al., 2013; Moernaut et al., 2014, 2015). Sumner et al. (2013) outline four independent criteria for identifying an earthquake trigger for turbidites: 1) synchronous turbidites in multiple basins; 2) observing an identical number of turbidites above and below the confluence of submarine channels; 3) identifying a larger relative volume for turbidites compared with others deposited in the same setting; 4) a historical or observational record of an earthquake that is coeval with the turbidite.

Nazaré canyon has three branching channels, with confluences at 500 and 3,000 m water depth (Fig. 2) (Lastras et al., 2009). There are no sediment cores positioned directly below or above these confluences, making the confluence test of Adams (1990) unsuitable in this study area. Due to a lack of core coverage in the basin it is also impossible to estimate turbidite volumes and invoke a seismic trigger. Turbidite paleoseismology has been previously applied to the Iberian margin and a catalogue of Late Quaternary seismo-turbidites exists for the Tagus and Horseshoe Basins (Fig. 10A) (Garcia-Orellana et al., 2006; Gràcia et al., 2010; Masson et al., 2011b). This serves as a basis to test for earthquake triggering of turbidites in the Iberian Abyssal Plain.

Comparison with published Iberian turbidite records reveals that only three turbidites correlate well into the adjacent Tagus Abyssal Plain (TAP), and only one turbidite is well correlated in all three basins (Fig. 10A). Other large turbidites from the Iberian Abyssal Plain (IAP) are not synchronous with proposed seismo-turbidites from the Tagus or Horseshoe Abyssal Plain. The Iberian Margin contains multiple faults as a result of the compressional rotation associated with the Azores-Gibraltar fracture zone to the south (Fig. 10B) (Buforn et al., 1988; Borges et al., 2001; Zitellini et al., 2004; Custódio et al., 2015). The Nazaré Fault (NF) and the Lower Tagus Valley Fault (LTVF) are quite proximal, suggesting that large earthquakes originating from the LTVF might be capable of triggering sediment failures which ignite and flush Nazaré Canyon (Fig. 10B) (Johnston, 1996). This could explain why several turbidites are synchronous in both the Tagus and Iberian Abyssal Plains (Fig. 10A).

The tectonic regime along the south-western section of the Iberian Margin is more complicated, with several extensive offshore faults capable of generating large earthquakes (Fig. 10B) (Pro et al., 2013; Custódio et al., 2015). There is 300 km distance from the south-west margin to Nazaré Canyon, and there is no continuous fault system between the areas. Paleoseismic intensity reconstructions and seismic propagation models indicate large earthquakes along the south-western section of the margin are unlikely to trigger canyon-flushing in Nazaré Canyon (Dobrovolsky et al., 1979; Buforn et al., 1988; Fukushima and Tanaka, 1990; Johnston, 1996; Zitellini et al., 2004). This could explain why several turbidites present in the Tagus and Horseshoe Abyssal Plain are not present in the Iberian Abyssal Plain (Fig 10A).

The synchronous deposition test does not strongly support regional earthquakes as a trigger for canyon-flushing turbidites in the Iberian Abyssal Plain. The historical record of large Portuguese earthquakes extends back to the Portugal and Galicia Earthquake at 60 BC (Galbis, 1932; Baptista and Miranda, 2009). Several earthquakes dating further back to 7,000 Cal years BP have been inferred from onshore paleo-tsunami records (Ruiz et al., 2008). Importantly, these records place the epicentres of many of these past earthquakes along the South-western Iberian Margin, and not proximal to Nazaré Canyon. This also makes them unsuitable as an independent earthquake archive with which to compare Iberian Abyssal Plain turbidite ages.

The Nazaré Fault (NF) forms part of a larger NNE-SSW-trending Variscan fault system which extends across the peninsula (Fig. 10B) (Buforn et al., 1988; Zitellini et al., 2004). Paleoseismic reconstructions on these NNE-SSW trending fault zones in Central and Northern Portugal are sparse, and no such record exists for the Nazaré Fault. This makes identifying possible earthquake triggering of canyon-flushing from nearby faults problematic. Optically Stimulated Luminescence (OSL) dates from fault surfaces along the Manteigas-Bragança Fault (MBF) north-east of the Nazaré Fault reveal three large (Mw 7.3) earthquakes between 11.5 and 14.5 ±2 ka, (Fig. 10A). These ages do not correspond well with turbidites in the Iberian Abyssal Plain, although the wide margins of error do encompass 2 turbidites at 11.7 ka and 14.9 ka (Fig. 10A). It has been proposed that these NNE-SSW-trending Variscan faults primarily accommodate reverse faulting and not strike-slip faulting. In contrast, WNW–ESE-trending faults on the South-western Iberian Margin exhibit strike-slip behaviour (Borges et al., 2001; Custódio et al., 2015). This would imply that unlike the southwestern margin, earthquakes along the Central Portuguese Margin do translate down-fault, limiting seismic propagation (Sylvester, 1988).

A small number of turbidites from the Iberian Abyssal Plain can be correlated with seismo-turbidites previously identified in the Tagus and Horseshoe Abyssal Plains (Fig. 10A). Turbidites which do not correlate in the Tagus or Horseshoe Basins could be the result of earthquakes originating from further north along the ENE-SSW Fault zone. However, given the wide uncertainties on earthquake ages, and the structure of the fault system, any correlation with canyon-flushing events would be largely speculative. There are multiple other intraplate faults that may trigger canyon-flushing, but without accurate paleoseismic records it is impossible to imply causation (Buforn et al., 1988; Zitellini et al., 2004; Villamor et al., 2012; Custódio et al., 2015). The lack of convincing evidence for earthquake triggered canyon-flushing along this margin highlights the problems of applying turbidite paleoseismology methods to structurally complex margins.

5.3 Climate change and geohazard implications

As previously outlined, many authors have suggested that large continental margin failures are associated with times of significant climatic and sea level change (Maslin et al., 2004; Owen et al., 2007; Lee, 2009; Smith et al., 2013; Brothers et al., 2013). One implication of this is that future climate and sea level change might increase the frequency of large continental margin failures. Our findings contribute to a growing body of literature that suggests large sediment failures in multiple settings are temporally random and are not significantly influenced by climate or sea change (Beattie and Dade, 1999; Urlaub et al., 2013, 2014; Clare et al., 2014; Hunt et al., 2014; Moernaut et al., 2014; Talling et al., 2014). It might be reasonable to conclude that the recurrence intervals of large sediment failures and canyon-flushing along the Central Iberian Margin will not be significantly influenced by any future sea level rise. Canyon-flushing in other deep-sea canyons is currently poorly understood and few estimates of recurrence rates exist. Additional work into other canyon systems would help to determine if temporal randomness is a wide-spread characteristic of canyon-flushing.

Our statistical analysis demonstrates that sea level is a significant control on canyon-filling. However, the current sea level highstand features limited canyon-filling events. These are typically small, channel-confined turbidity currents that die before reaching the lower canyon (Fig 9A). Intuitively, we might expect future sea level rise to further limit canyon-filling turbidity currents, as past sea level rise did. However, future sea level rise has been predicted to increase rates of coastal erosion (Gornitz, 1991; Bray and Hooke, 1997; Leatherman et al., 2000). This increase in sediment delivery to the Iberian Shelf could lead to an increase in littoral transport into the head of Nazaré Canyon, thereby promoting greater instability. These smaller turbidity currents may not have large geohazard implications, but are important for assessing risk to seafloor structures such as pipelines and telecommunication cables.

6. Conclusions

Canyon-flushing turbidity currents have been predicted to occur much less frequently than those that fill canyons (Paull et al., 2005; Canals et al., 2006; Arzola et al., 2008; Puig, et al., 2014). Here we demonstrate that canyon-flushing turbidity currents in this system have recurrence rates of several thousands of years on average. This recurrence rate is an order of magnitude longer than those filling the canyon during sea level lowstand, and more than two orders of magnitude longer than those that fill the canyon in the present day. The recurrence intervals for canyon-filling and canyon-flushing appear to have distinctly different statistical distributions. Moreover, unlike canyon filling, the recurrence rate of canyon-flushing events does not appear to be affected by long-term changes in sea level. This suggests that they may have different triggers, or that the signal of triggering mechanisms is shredded due to the long recurrence intervals. From a geohazard assessment perspective, this implies that the frequency of potentially hazardous canyon-flushing events may not be influenced by future sea level predictions. The tectonic complexity of the margin, uncertainties in age-control, and the time-independent behaviour of canyon-flushing make determining a trigger for canyon-flushing problematic; although earthquakes may trigger some events.

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Figures and Tables:

Fig. 1: Map of the Portuguese Margin, showing the location of the main sedimentary basins and their feeder canyons. Core JC27-51, located in the distal Iberian Abyssal Plain is also shown. Bathymetry data are from the GEBCO database (IOC, IHO, BODC, 2003).

Fig. 2: Location map of the Nazaré Canyon and the locations of canyon cores used in this study.

Fig. 3: Lithological description of the cores used in this study. Dark grey represents turbidite mud, orange represent silt and yellow represents sand. Radiocarbon dates and their core positions are indicated with black arrows.

Fig. 4: Bed types present in JC27 sediment cores. JC27-51: (A) Thick mud-dominated turbidites wich bioturbated mud caps, (B) thick sand rich turbidites with discernible Bouma (T) divisions , and (C) a debrite with mud and silt clasts. JC27-46: (D) The upper 6 m of the core is composed of thin sand or silt-based turbidite with a dominant mud unit and hemipelagic sediment intervals. (E) The lower 6m is mainly composed of thin-bedded sand rich turbidites with a thinner mud cap and sparse hemipelgic sediment. JC27-47: Turbidites are typically (F) planar and often erosive, or (G) chaotically structured, with reverse grading, overturned layers and evidence of scouring.

Fig. 5: Hemipelagic material and turbidites present within cores JC27-51 and JC27-46. ITRAX Calcium records assist in defining the boundaries between pure hemipelagite and bioturbated turbidite mud cap in JC27-51. There are no ITRAX data for JC27-46, but hemipelagite can still be defined by the presence of a colour change and dark mottling.

Fig. 6: hemipelagic aGe models for cores JC27-51 and JC27-46. R^2 values indicate minimal deviation from the line and thus a relatively stable hemipelagic sedimentation rate in both cores. This serves as a reasonable basis for using a stable sedimentation rate to interpolate between radiocarbon dates and further back in the record beyond radiocarbon age.

Fig. 7. **A**: Frequency of turbidites contained within core external levee JC27-46. Turbidites are binned into 500 year intervals and plotted against global eustatic sea level curves. Trend suggests near

shutdown of large canyon-filling turbidites recorded in the levee following the onset of sea level highstand at 7 ka. **B**: Frequency of turbidites from internal levee core JC27-47. It illustrates a similar pattern of decline through sea level transgression and into the present-day highstand. H1-3 = Heinrich events. YD = Younger Dryas.

Fig. 8: Comparison of Iberian Abyssal Plain recurrence intervals with published recurrence intervals from several long-term basin records in Clare et al. (2014). The vertical axis plots the probability (P) that a given turbidite recurrence interval value (time since last event, T) will exceed the average recurrence interval for its respective dataset. The horizontal axis plots normalised recurrence interval data. The recurrence intervals are normalized by subdividing each recurrence interval (T) by the mean recurrence interval (λ) for each of the data sets to plot a dimensionless variable, R_T. N = number of events in each basin data series.

Fig. 9: Schematic displaying the spatial and temporal variability in turbidity current frequency in Nazaré Canyon during periods of sea level lowstand and highstand. **A**: During the present day highstand canyon-filling turbidity currents (TCs) are frequent in the upper canyon, but only small thalweg-confined turbidity currents reach to greater than 4,000 m water depth. **B**: During lowstand conditions canyon-filling turbidity currents were much larger, were able to reach the lower canyon, and regularly over-spilled canyon levees. Infrequent canyon-flushing turbidity currents were not affected by changes in sea level and continued into the present-day highstand.

Fig. 10. **A**: Proposed seismoturbidites from the Tagus and Horseshoe Abyssal Plains from Grácia et al. (2010) and Masson et al. (2011b) (shown as black dots). Turbidites from the Iberian Abyssal Plain are shown in red, and paleoearthquakes from the Manteigas-Braçanga Fault are shown in blue. Orange bars represent event which can be correlated based on their age uncertainties. Dashed error bars on ages represent ages that are determined through linear interpolation. **B**: Map of the Portuguese Margin showing primary structural faults. MTB = Manteigas-Braçanga Fault, NF = Nazaré Fault, LTVF

= Lower Tagus Valley Fault, MF = Messejana Fault, PSF = Pereira de Sousa Fault, MPTF - Marquês de Pombal Thrust Fault, HTF = Horseshoe Thrust Fault, GTF = Gorringe Thrust Fault, SWIM = South-western Iberian Margin lineaments. Labels in italics indicate locations of paleoseismic reconstructions in Fig. 10A.

Table 1: List of radiocarbon samples used in this study. Note: Cal BP ages from D15738 and D15739 (Masson et al, 2011a) are median values, and not determined using maximum probability.

Table 2: Core JC27-51 turbidites, their projected ages, and recurrence interval determined using a linear age model.

Supplementary table 1: Hemipelagic thicknesses, hemipelagic linear age model and interpolated ages for turbidites in levee core JC27-46.

Supplementary table 2: Model outputs from both the generalised linear models and Cox proportional hazards models testing the influence of sea level on turbidite recurrence and frequency in levee core JC27-46. Asterisks represent the significance of the p-values, with '***' being the highest significance, and no asterisks indicating no significance of sea level as an explanatory variable. Q-Q plot fits are indicated as Very poor, Poor, Moderate, or Good. The results indicate that sea level is a significant explanatory variable for changes in turbidite recurrence and frequency.

Supplementary table 3 Model output for Mann-Whitney and Kolgomorov-Smirnoff tests of the distribution of Iberian Abyssal Plain turbidite recurrence in core JC27-51 vs. those of other basin plains and levee core JC27-46. The results indicate that the distribution of recurrence intervals in the Iberian Abyssal Plain cannot be distinguished from those of other basin plains that exhibit a Poisson distribution. The results also indicate that the Iberian Abyssal Plain also has a different distribution of recurrence intervals from levee core JC27-46, suggesting a different long-term control on recurrence.

Figure 1

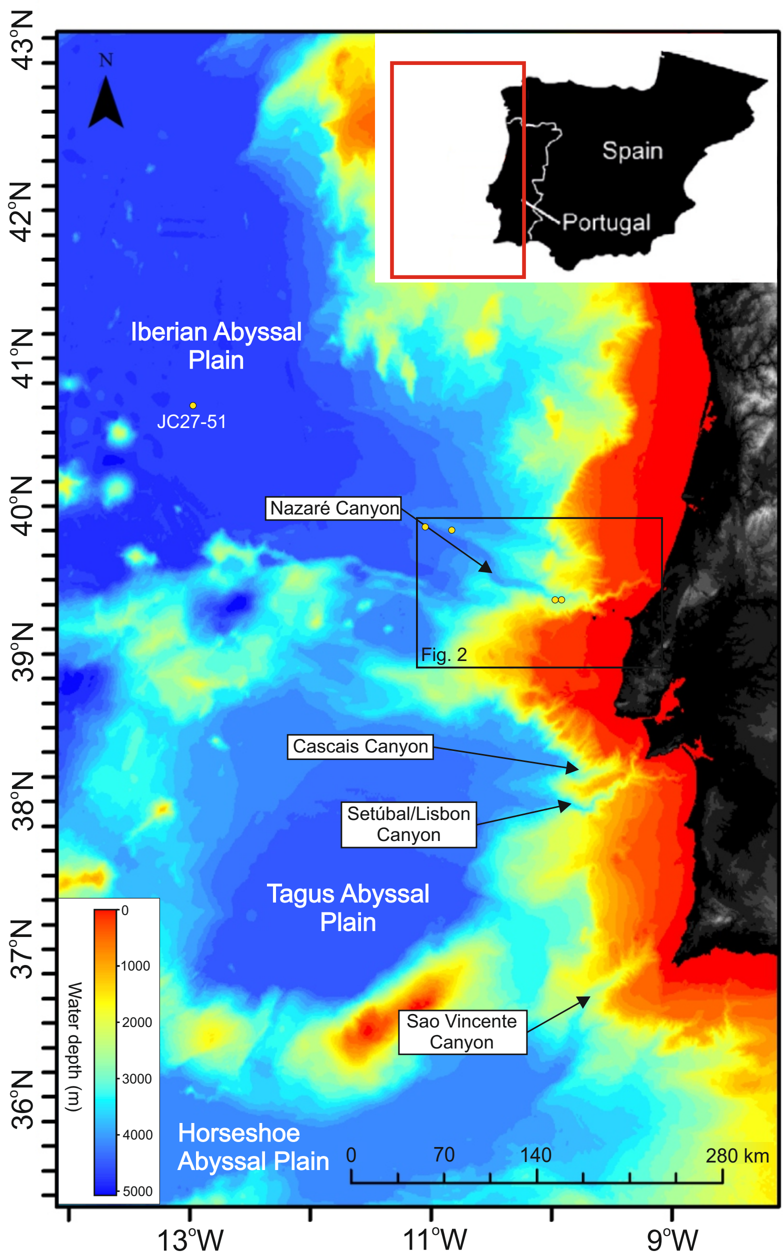
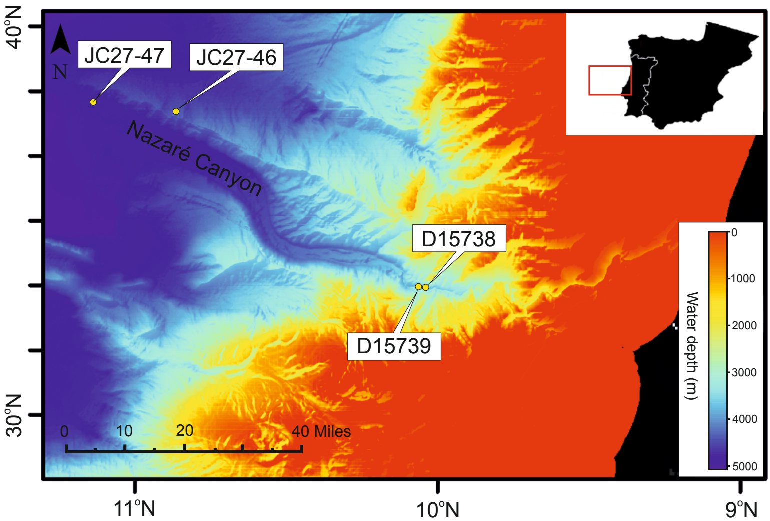

43°N
42°N
41°N
40°N
39°N
38°N
37°N
36°N
13°W
11°W
9°W
N
Spain
Portugal
Iberian Abyssal Plain
JC27-51
Nazaré Canyon
Fig. 2
Cascais Canyon
Setúbal/Lisbon Canyon
Tagus Abyssal Plain
Sao Vincente Canyon
Horseshoe Abyssal Plain
Water depth (m)
0
1000
2000
3000
4000
5000
0
70
140
280 km

Figure 2



40°N
30°N
N
JC27-47
JC27-46
Nazaré Canyon
D15738
D15739
0
10
20
40 Miles
11°N
10°N
9°N
Water depth (m)
0
1000
2000
3000
4000
5000

Figure 3

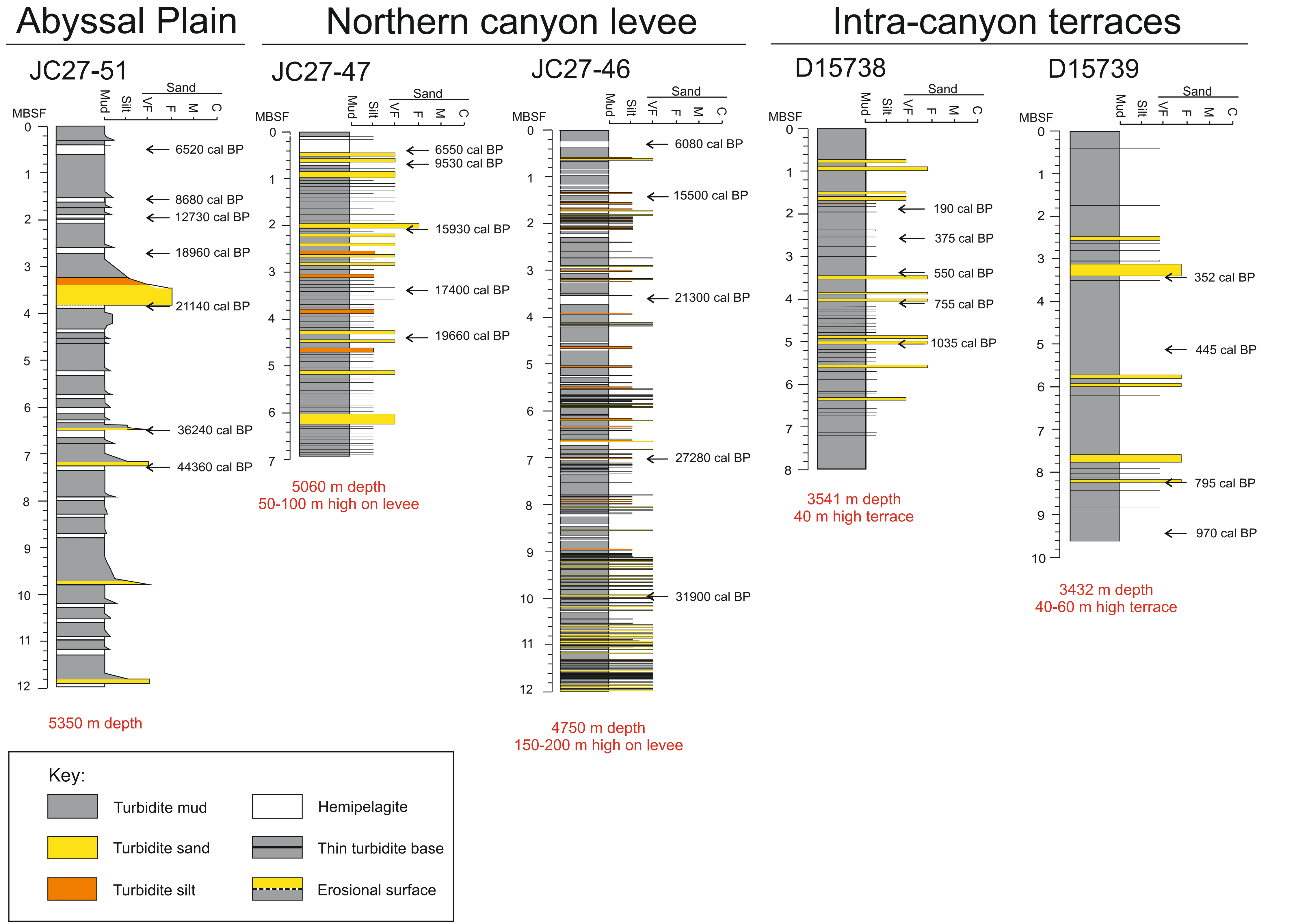

Abyssal Plain
Northern canyon levee
Intra-canyon terraces
JC27-51
JC27-47
JC27-46
D15738
D15739
MBSF
Mud
Silt
Sand
VF
F
M
C
6520 cal BP
8680 cal BP
12730 cal BP
18960 cal BP
21140 cal BP
36240 cal BP
44360 cal BP
5350 m depth
6550 cal BP
9530 cal BP
15930 cal BP
17400 cal BP
19660 cal BP
5060 m depth
50-100 m high on levee
6080 cal BP
15500 cal BP
21300 cal BP
27280 cal BP
31900 cal BP
4750 m depth
150-200 m high on levee
190 cal BP
375 cal BP
550 cal BP
755 cal BP
1035 cal BP
3541 m depth
40 m high terrace
352 cal BP
445 cal BP
795 cal BP
970 cal BP
3432 m depth
40-60 m high terrace
Key:
Turbidite mud
Hemipelagite
Turbidite sand
Thin turbidite base
Turbidite silt
Erosional surface

Figure 4

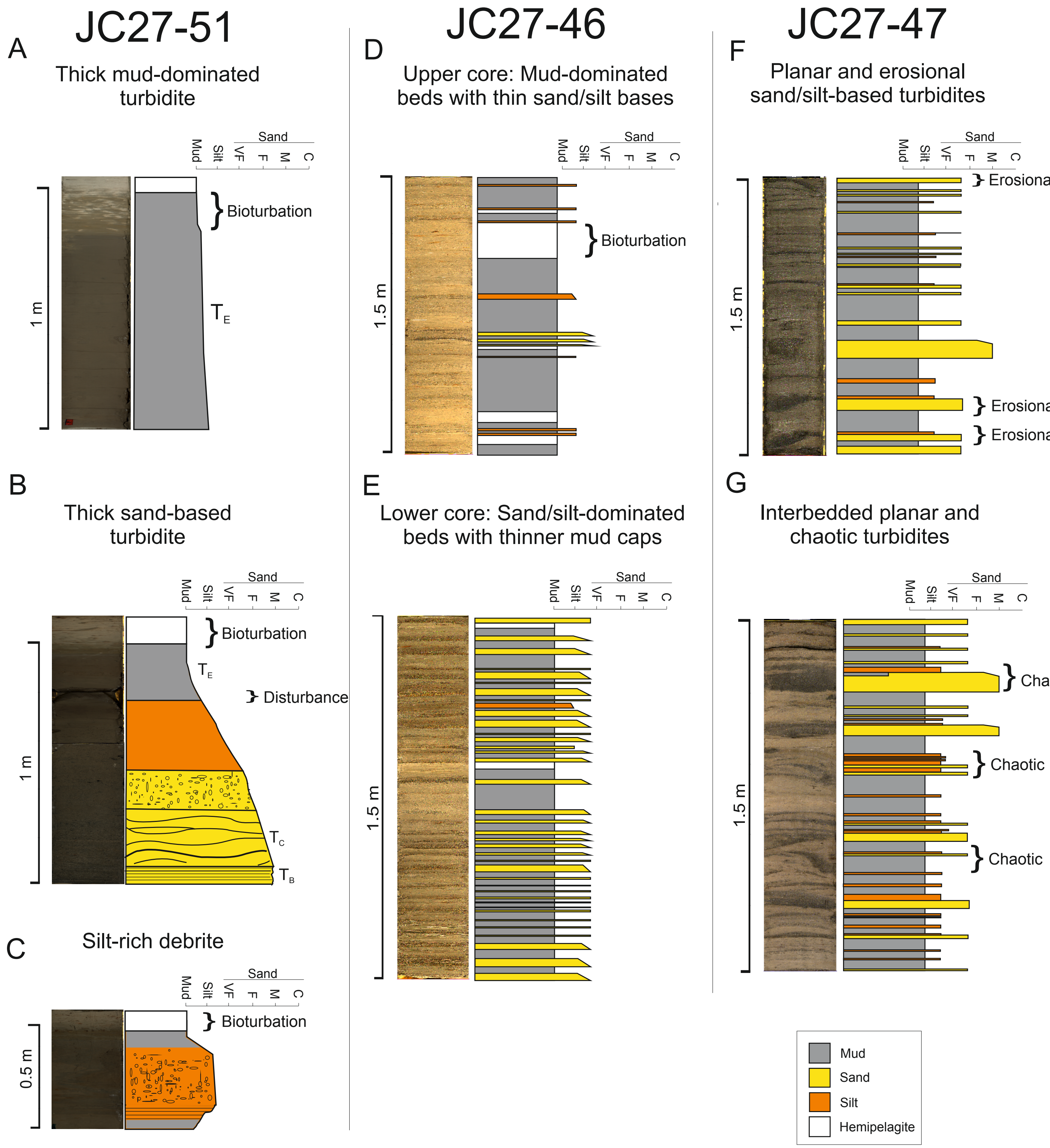

JC27-51
A
Thick mud-dominated turbidite
Mud
Silt
Sand
VF
F
M
C
Bioturbation
1 m
T E
B
Thick sand-based turbidite
Bioturbation
T E
Disturbance
T C
T B
C
Silt-rich debrite
Bioturbation
0.5 m
JC27-46
D
Upper core: Mud-dominated beds with thin sand/silt bases
1.5 m
E
Lower core: Sand/silt-dominated beds with thinner mud caps
JC27-47
F
Planar and erosional sand/silt-based turbidites
Erosiona
G
Interbedded planar and chaotic turbidites
Cha
Chaotic
Mud
Sand
Silt
Hemipelagite

Figure 5

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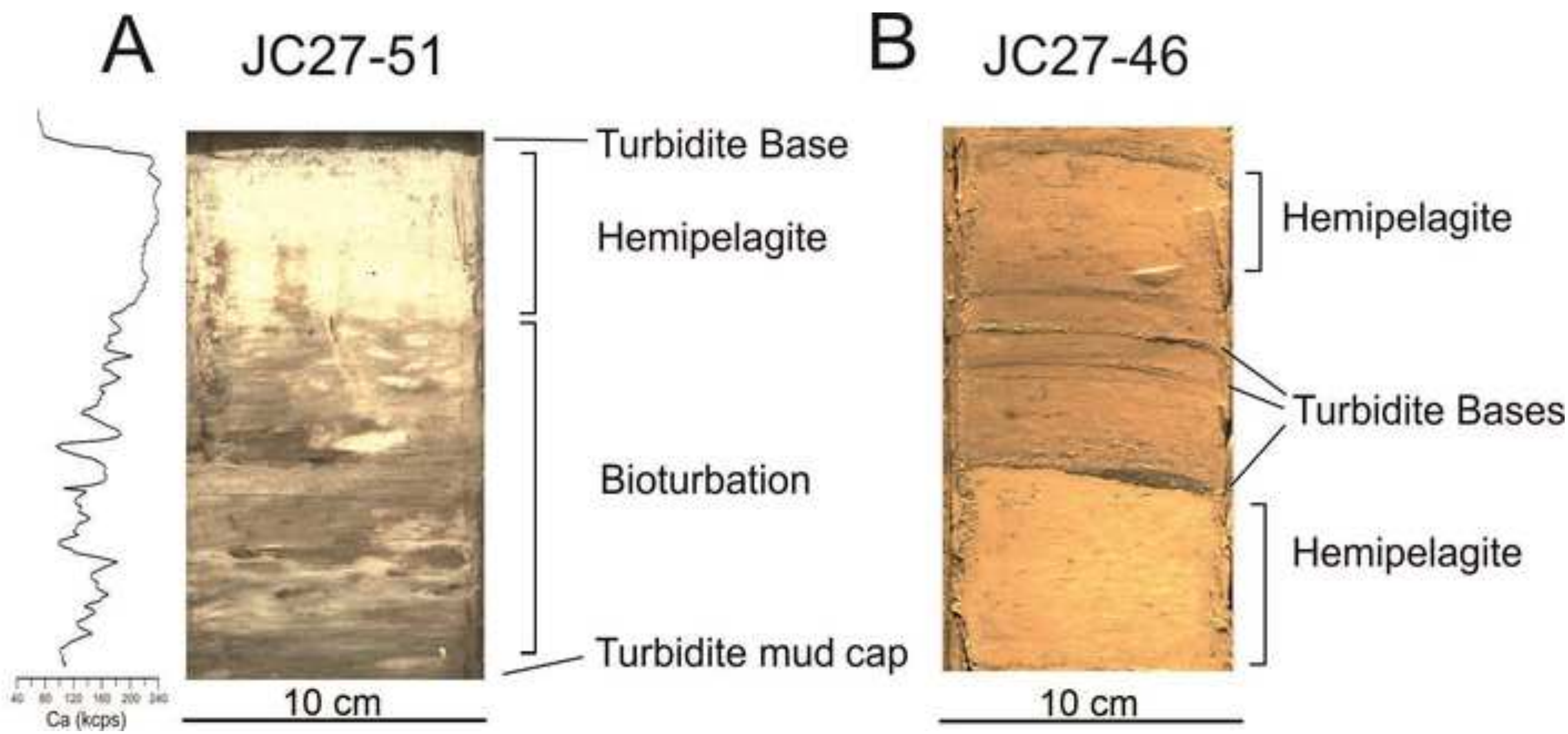
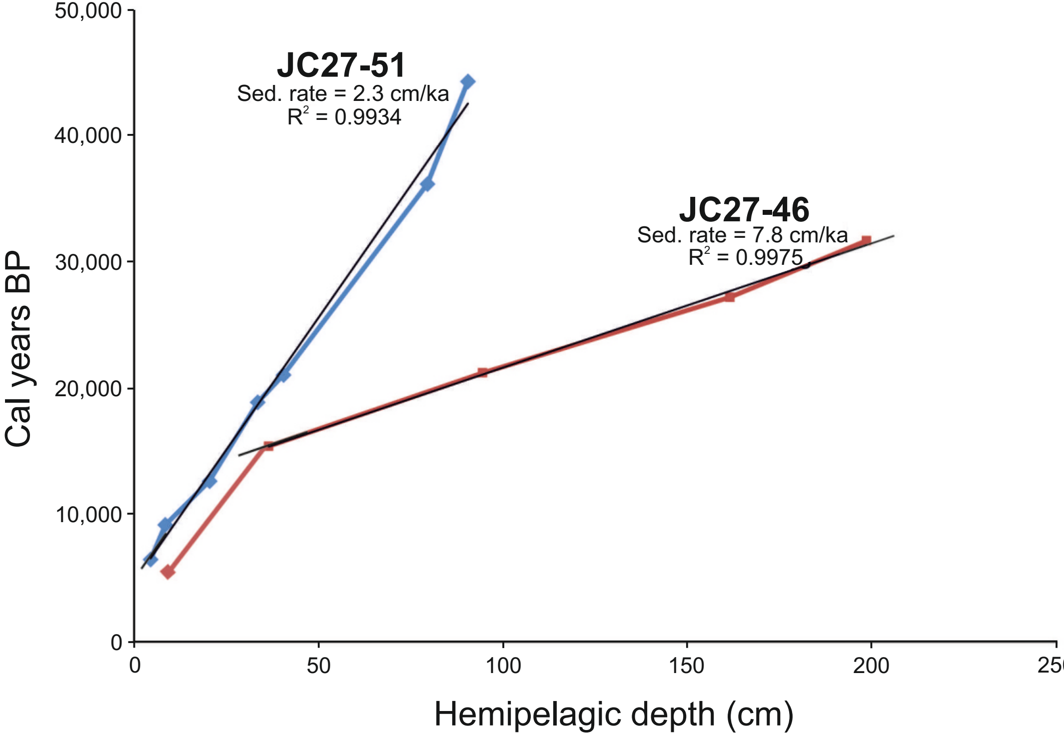
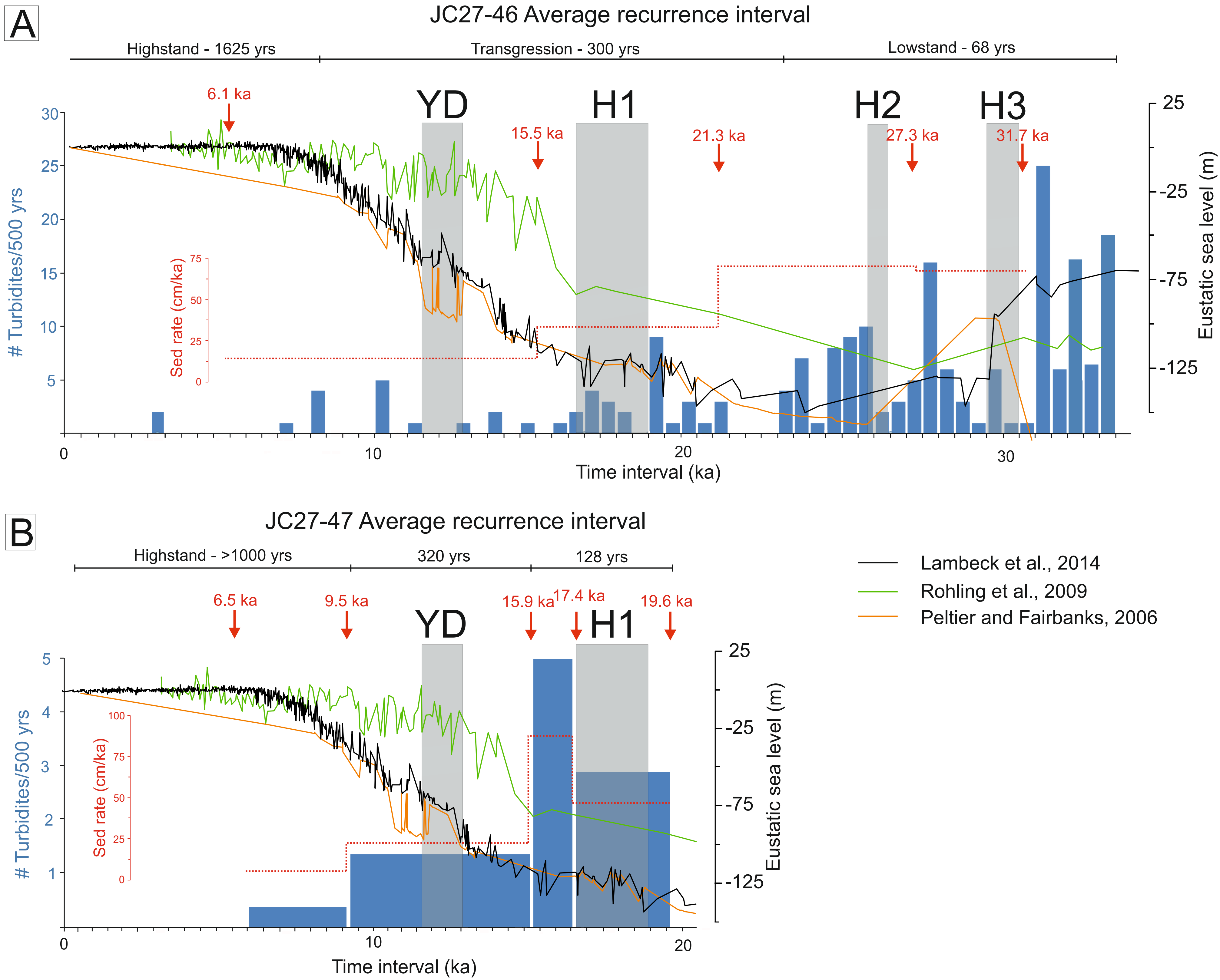


Figure 6



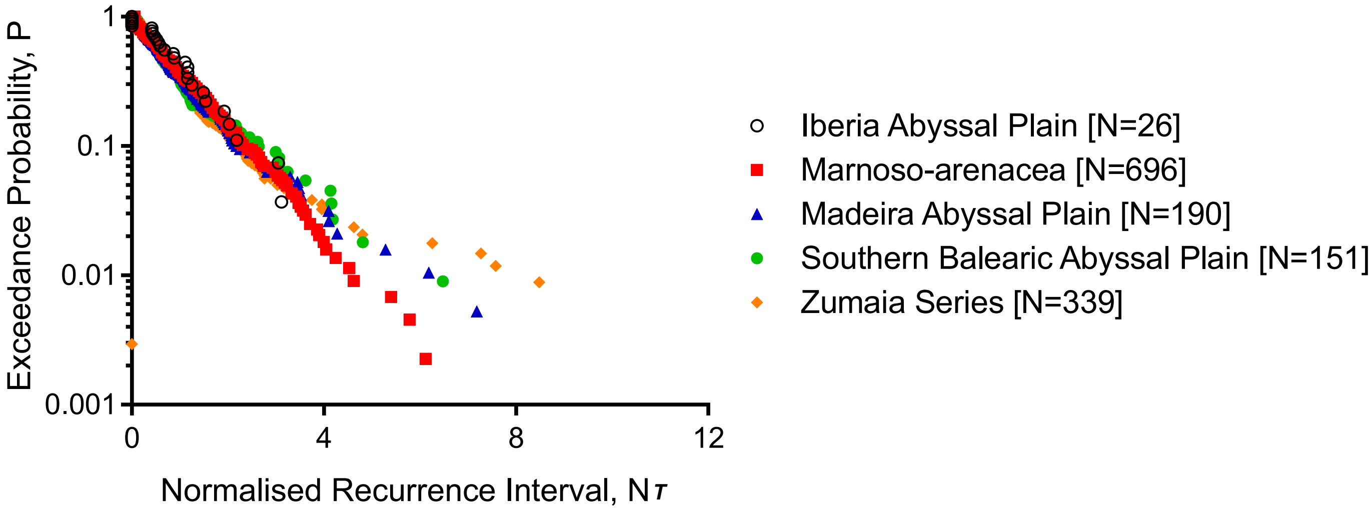
JC27-51
Sed. rate = 2.3 cm/ka
R^2 = 0.9934
JC27-46
Sed. rate = 7.8 cm/ka
R^2 = 0.9975
Cal years BP
50,000
40,000
30,000
20,000
10,000
0
0
50
100
150
200
250
Hemipelagic depth (cm)

Figure 7



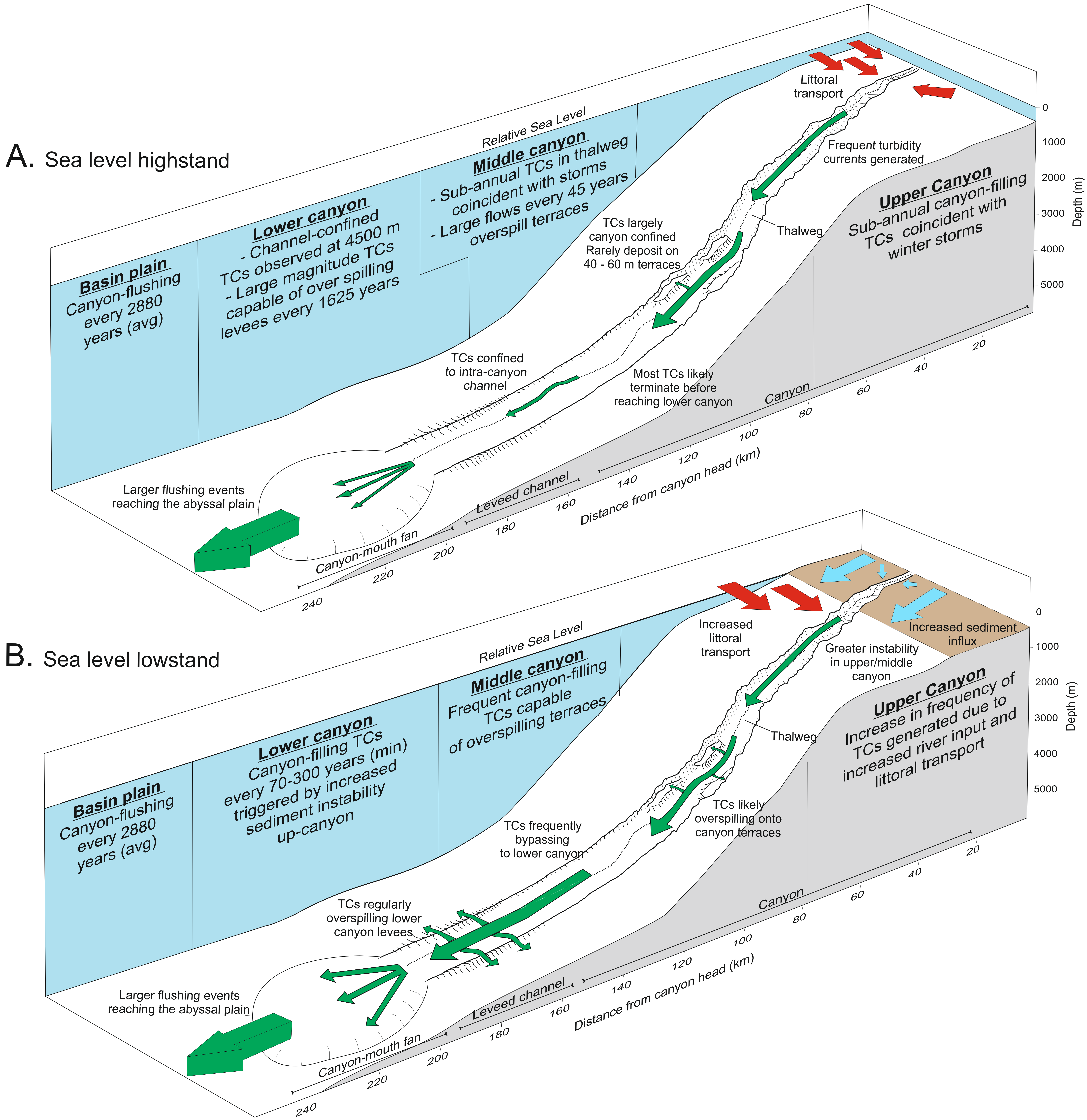
A
JC27-46 Average recurrence interval
Highstand - 1625 yrs
Transgression - 300 yrs
Lowstand - 68 yrs
YD
H1
H2
H3
6.1 ka
15.5 ka
21.3 ka
27.3 ka
31.7 ka
Turbidites/500 yrs
Sed rate (cm/ka)
Eustatic sea level (m)
Time interval (ka)
B
JC27-47 Average recurrence interval
Highstand - >1000 yrs
320 yrs
128 yrs
6.5 ka
9.5 ka
15.9 ka
17.4 ka
19.6 ka
Lambeck et al., 2014
Rohling et al., 2009
Peltier and Fairbanks, 2006

Figure 8



1
0.1
0.01
0.001
Exceedance Probability, P
0
4
8
12
Normalised Recurrence Interval, Nτ
Iberia Abyssal Plain [N=26]
Marnoso-arenacea [N=696]
Madeira Abyssal Plain [N=190]
Southern Balearic Abyssal Plain [N=151]
Zumaia Series [N=339]

Figure 9



A. Sea level highstand
Relative Sea Level
Littoral transport
Frequent turbidity currents generated
Upper Canyon
Sub-annual canyon-filling TCs coincident with winter storms
Middle canyon
- Sub-annual TCs in thalweg coincident with storms
- Large flows every 45 years overspill terraces
Lower canyon
- Channel-confined TCs observed at 4500 m
- Large magnitude TCs capable of over spilling levees every 1625 years
Basin plain
Canyon-flushing every 2880 years (avg)
TCs largely canyon confined Rarely deposit on 40 - 60 m terraces
Thalweg
TCs confined to intra-canyon channel
Most TCs likely terminate before reaching lower canyon
Larger flushing events reaching the abyssal plain
Canyon-mouth fan
Leveed channel
Canyon
Distance from canyon head (km)
Depth (m)
B. Sea level lowstand
Relative Sea Level
Increased littoral transport
Increased sediment influx
Greater instability in upper/middle canyon
Upper Canyon
Increase in frequency of TCs generated due to increased river input and littoral transport
Middle canyon
Frequent canyon-filling TCs capable of overspilling terraces
Lower canyon
Canyon-filling TCs every 70-300 years (min) triggered by increased sediment instability up-canyon
Basin plain
Canyon-flushing every 2880 years (avg)
Thalweg
TCs likely overspilling onto canyon terraces
TCs frequently bypassing to lower canyon
TCs regularly overspilling lower canyon levees
Larger flushing events reaching the abyssal plain
Canyon-mouth fan
Leveed channel
Canyon
Distance from canyon head (km)
Depth (m)

Figure 10

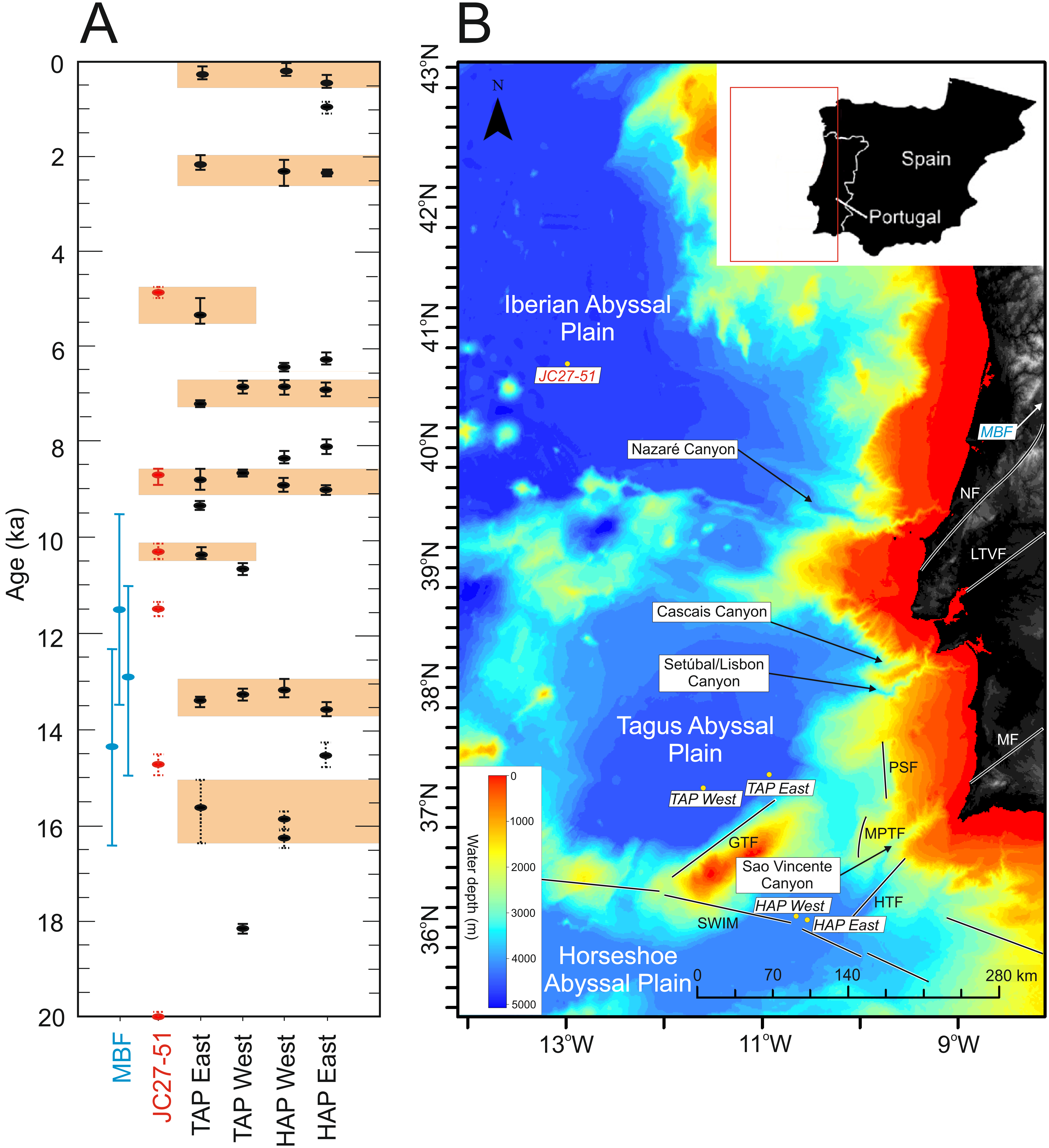

A
B
Age (ka)
0
2
4
6
8
10
12
14
16
18
20
MBF
JC27-51
TAP East
TAP West
HAP West
HAP East
43°N
42°N
41°N
40°N
39°N
38°N
37°N
36°N
13°W
11°W
9°W
N
Spain
Portugal
Iberian Abyssal Plain
JC27-51
Nazaré Canyon
MBF
NF
LTVF
Cascais Canyon
Setúbal/Lisbon Canyon
Tagus Abyssal Plain
MF
PSF
TAP East
TAP West
MPTF
GTF
Sao Vincente Canyon
HAP West
HTF
HAP East
SWIM
Horseshoe Abyssal Plain
Water depth (m)
0
1000
2000
3000
4000
5000
0
70
140
280 km

Table 1
Click here to download Table: Table 1.xlsx

Lab code	Core #	Core depth (cm)	Conventional age (BP)	Max. probabilty (Cal BP)	1σ Cal age ranges (Cal BP)	2σ Cal age ranges (Cal BP)
BETA-385402	JC27-46	32-33	5,940 ± 30	6,078	5,995-6,165	5,918-6,233
BETA_385403	JC27-46	158-159	13,639 ± 50	15,494	15,334-15,629	15,234-15,768
BETA_385404	JC27-46	349-350	18,290 ± 60	21,303	21,143-21,465	20,994-21,599
BETA_385405	JC27-46	701-702	23,600 ± 100	27,280	27,163-27,406	27,020-27,514
BETA_385406	JC27-46	996-997	28,790 ± 140	31,750	31,568-32,146	31,449-32,521
SUERC-31798	JC27-47	53-54	6,120 ± 35	6,270	6,197-6,341	6,117-6,422
BETA-401321	JC27-47	70.5-71.5	8,510 ± 30	8,760	8,652-8,889	8,584-8,972
SUERC-31799	JC27-47	212-214	13,530 ± 47	15,300	15,201-15,479	15,117-15,638
BETA-401322	JC27-47	340-341.5	14,910 ± 50	14,700	17,211-17,469	17,089-17,576
SUERC-31802	JC27-47	445-448	16,910 ± 55	19,580	19,484-19,725	19,360-19,894
	JC27-51	48.5-49.5	6,355 ± 37	6,520	6,433-6,604	6,342-6,680
BETA-382057	JC27-51	155-156	8,470 ± 30	8,680	8,604-8,769	8,567-8,893
	JC27-51	196.5-197.5	11,535 ± 39	12,730	12,661-12,810	12,613-12,904
	JC27-51	264-265	16,415 ± 53	18,960	18,883-19,083	18,808-19,206
	JC27-51	381-382	18,196 ± 62	21,140	21,019-21,322	20,886-21,469
BETA-382053	JC27-51	649-650	33,010 ± 200	36,240	36,000-36,454	35,722-36,719
BETA-382054	JC27-51	726-727	41,500 ± 550	44,360	43,825-44,875	43,310-45,314
SUERC-18143	D15738	178.5	573 ± 37	190		90-290
SUERC_18146	D15738	250.5	738 ± 37	375		290-460
SUERC_18147	D15738	335	934 ± 37	550		480-620
SUERC_18148	D15738	405	1,204 ± 37	755		660-850
SUERC_18149	D15738	509.5	1,482 ± 35	1,035		940-1,130
SUERC-18150	D15739	334.5	703 ± 37	352		270-435
SUERC-18151	D15739	489.5	821 ± 37	445		370-520
SUERC-18153	D15739	829.5	1,245 ± 37	795		700-890
SUERC-18156	D15739	959.5	1,411 ± 37	970		890-1,050

Table 2
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Turbidite no	Turbidite base depth (cm)	Turbidite thickness (cm)	Approx age (cal BP)	Recurrence interval (years)
1	25	25	4,850	0
2	45	20	4,850	4,430
3	155	103	8,680	1,700
4	172	13	10,300	0
5	187	15	10,300	1,200
6	194	4	11,505	3,200
7	254	53	14,700	5,550
8	378	110	20,250	1,350
9	428	46	21,600	1,500
10	454	22	23,100	0
11	468	14	23,100	0
12	503	35	23,100	1,600
13	523	16	24,700	2,550
14	570	40.5	27,250	1,950
15	597	24	29,200	4,300
16	618	10	33,500	1,200
17	645	24	34,700	9,000
18	671	12.5	43,700	0
19	725	54	43,700	3,600
20	790.5	57.5	47,300	5,850
21	837	36	53,150	1,250
22	870	30.5	54,400	8,800
23	978	102	63,200	3,350
24	1016	38	66,550	3,350
25	1050	26	69,900	6,300
26	1091	36	76,200	2,500
27	1119	20	78,700	3,350
28	1187	60	82,050	
				Avg: 2800

Supplementary table 1

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Supplementary table 2

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Supplementary table 3

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