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Fluvial-system response to climate change: the Paleocene-Eocene Tremp Group, Pyrenees, Spain

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Abstract

The Tremp Group of the Tremp-Graus Basin (Southern Pyrenees, Spain) is a succession of predominantly continental origin that records the Paleocene-Eocene Thermal Maximum (PETM), a transient episode of extreme global warming that occurred across the Paleocene-Eocene boundary. For this succession, the stratigraphic position of the PETM is accurately determined, and histories of tectonic and sea-level controls are well constrained. Building upon previous studies, this work assesses changes in sedimentary architecture through the PETM in the Tremp Group, based on quantitative sedimentological analyses documented over a km-scale strike-oriented transect in the Arén area, with the scope to better understand the response of this alluvial system to the hyperthermal event. The analysed features represent a partial record of the geomorphic organization and processes of the system at the time of deposition, and are therefore interpretable in terms of geomorphic change in alluvial landscapes caused by the PETM.

The record of the PETM, as previously recognized, begins at a time when erosional palaeotopographic relief was developed and deposition was confined in valleys. A shift between valley back-filling and widespread aggradation is observed at the onset of the PETM interval, which demonstrates uniquely the impact of the hyperthermal on both depositional loci and interfluves. Compared to underlying strata, the interval that embodies the onset and main phase of the PETM is characterized by: (i) higher proportion of channel deposits; (ii) channel complexes of greater average thickness and width; (iii) barforms and channel fills that are slightly thicker; (iv) increased thickness of sets of cross-stratified sandstones; (v) similar values of maximum extraclast size, by architectural element. An evident change in the facies organization of channel deposits is also seen through the stratigraphy, though this appears to predate the PETM.

Increased channel-body density in the PETM interval can be explained in terms of increased channel mobility, which itself can be related to changes in the stream catchments (e.g., greater bedload delivery, increased water discharge or discharge variability), or to changes in the nature of the depositional basin that would permit the channels to be more mobile (e.g., increased bank erodibility due to variations in vegetation type and density). Interfluve planation is inferred to have occurred immediately prior to, or penecontemporaneously with, accumulation of PETM deposits, which is in accord with inferences of increased erodibility of the interfluves or increased stream erosive power. These observations offer insight into the potential geomorphic metamorphosis of river systems in mid-latitude regions experiencing conditions of rapid global warming.

Keywords: Paleocene-Eocene Thermal Maximum; PETM; fluvial channel; alluvial; fluvial architecture; global warming.

Introduction

A brief episode of extreme global warming related to the release of isotopically light carbon into the Earth’s oceans and atmosphere is recorded across the Paleocene-Eocene boundary (ca. 56 Ma; Kennett & Stott 1991; Koch et al. 1992; Charles et al., 2011). This event is variably referred to as Initial Eocene Thermal Maximum, Latest Paleocene Thermal Maximum, or Paleocene-Eocene Thermal Maximum (PETM), and is believed to have been characterized by a rise in global temperatures of 5º to 9º C in 10 kyr, followed by a gradual decline to pre-PETM values that took between 100 and 200 kyr (Kennett & Stott 1991; Zachos et al. 2003; 2006; Tripati & Elderfield 2005; Sluijs et al. 2006; McInerney & Wing 2011). The stratigraphic expression of the PETM is often recognized by a characteristic negative shift in $\delta^{13}C$ (carbon isotope excursion; CIE), detected in both marine and terrestrial strata at the Paleocene-Eocene boundary (Kennett & Stott

1991; Koch et al. 1992; McInerney & Wing 2011, and references therein), although recent evidence suggests that a warming trend was already established before the time recorded in the CIE (Secord et al. 2010; Bowen et al. 2015)

Recent research has concentrated on how the PETM global changes affected sediment routing systems, through controls on weathering (e.g., Bolle & Adatte 2001, and references therein; Dallanave et al. 2010; Dypvik et al. 2011), sediment-delivery processes and rates (e.g., Schmitz & Pujalte 2003; Minelli et al. 2013), and type, distribution and characteristics of sedimentary sinks (e.g., Schmitz & Pujalte 2007; Foreman et al. 2012; Foreman 2014). Research effort has also been dedicated to assessing the impact of the PETM hyperthermal event on the geomorphic change of terrestrial landscapes, as documented in the stratigraphic record. In particular, recent publications provide insight into the sedimentary record of the PETM for some continental successions in which the stratigraphic position of the PETM is readily identifiable (Foreman et al. 2012; Foreman 2014; Kraus et al. 2015). At the PETM, continental environments may have variably been affected by the controls exerted by climate change on the eustatic level (Sluijs et al. 2008), on regimes of water and sediment discharge to the river systems, and on characteristics of catchments, floodplains, and interfluves, which are all themselves related to a number of climatically controlled variables.

A sedimentary succession of predominantly continental origin in which the stratigraphic position of the PETM-related CIE is well constrained is the Tremp Group of the Tremp-Graus Basin, Spanish Pyrenees (Schmitz & Pujalte 2003; 2007; Pujalte et al. 2014). Stratigraphic changes observed in this succession across the CIE have been interpreted in terms of responses of river systems to the PETM (Schmitz & Pujalte 2007).

Building upon previous studies, the aim of this work is to assess changes in sedimentary architecture through the PETM in the Tremp Group, based on results from original facies and architectural-element analyses, with the scope to better understand the response of the alluvial landscapes to the hyperthermal event.

The analysed features are considered to reflect – in part – the geomorphic organization and processes of the fluvial system at time of deposition. Lithofacies and associations thereof are indicative of different processes, including primary depositional processes (e.g., deposition in upper versus lower flow-regime conditions), and sub-environments (e.g., channel versus overbank), respectively. Furthermore, architectural features of sedimentary bodies, such as the geometry and nature of external and internal bounding surfaces and the relative spatial arrangement of these bodies, bear a record of the original geomorphology of the depositional systems. They relate landform types, their modes and rates of evolution, and their genetic relationships, although not always in a straightforward manner (cf. Bridge 2003; and references therein). On the basis of observations of the depositional architecture, inferences are here made about the geomorphic change in alluvial landscapes associated with the PETM.

The specific objectives of this study are: (i) to provide a quantitative description of changes in sedimentary architecture across the Paleocene-Eocene boundary in the Tremp Group, particularly as documented over a km-scale strike-oriented transect; (ii) to discuss the significance of these changes in terms of interpreted variations in geomorphic processes, landforms and associated drivers.

Geological setting

The Pyrenees are an asymmetrical chain of mountains composed of doubly vergent thrust wedges that formed in response to the collision between the Iberian and the European plates, during the late Cretaceous to early Miocene (Puigdefàbregas & Souquet 1986; Vergés et al. 1995). The south-central Pyrenees (Spain) consist of a fold-and-thrust belt divided into three imbricate west-trending thrust sheets, which are referred to as the Bóixols, Montsec and Sierras Marginales, from north to south, respectively (Burbank et al. 1992; figure 1A). Foreland basins developed on the front of these thin-skinned thrust sheets. In the early Paleogene, one of these basins, the Tremp-Graus Basin, evolved as a piggy-back basin (*sensu* Ori & Friend 1984) on top of the Montsec thrust sheet, south of the Bóixols thrust (Eichenseer & Luterbacher 1992; Puigdefàbregas et al. 1992; Luterbacher 1998). The timing of thrust activity is reflected in the southward piggy-back propagation and dictates depocentre migration. The Bóixols thrust sheet was emplaced during the Late Cretaceous (Berástegui et al. 1990; Bond & McClay 1995; Fernández et al. 2012). Before the Paleocene-Eocene boundary, sedimentation in the Tremp-Graus Basin was contemporaneous with the incipient propagation of the Montsec thrust, although most of the activity of this lineament took place in the early Eocene (Williams 1985; Farrell et al. 1987; Puigdefàbregas et al. 1992; Sinclair et al. 2005; Fernández et al. 2012). At these times, the northern part of the Tremp-Graus Basin was being affected by the activity of underlying blind thrusts (Eichenseer & Luterbacher 1992; Luterbacher 1998). The Montsec frontal ramp and Bóixols thrust likely determined a basin physiography characterized by a steeper northern margin and a gentler southern one. The basin displayed a WNW-oriented axis, parallel to the orogen, and the drainage was consistently towards the Bay of Biscay to the west (Plaziat 1981; Cuevas 1992; Vergés & Burbank 1996; Whitchurch et al. 2011; figure 1).

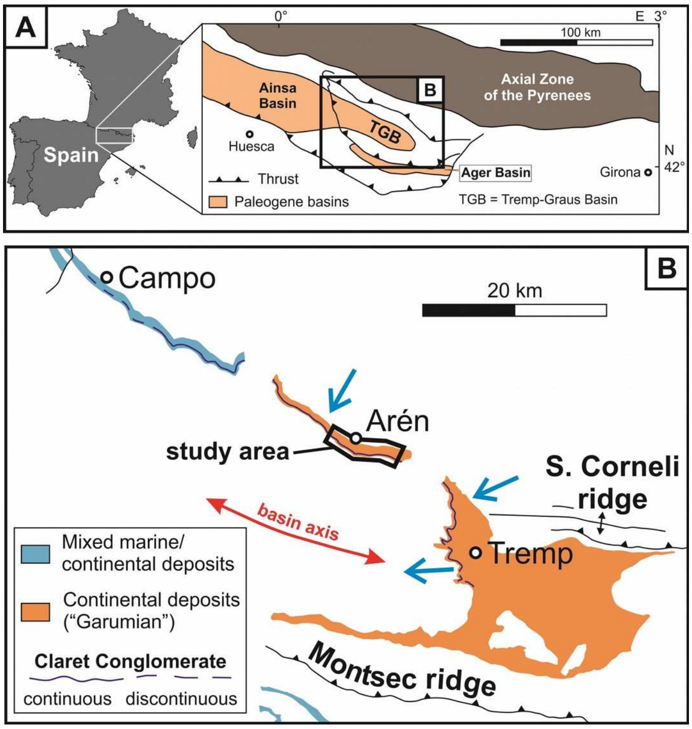


Figure 1: A) Location of the Tremp-Graus Basin in the structural setting of the southern Pyrenees. B) Location of the study area on outcrop map of the early Paleogene deposits of the Tremp-Graus Basin. The arrows indicate general drainage directions. The trace of the Claret Conglomerate marks the position of the PETM. Modified after Pujalte et al. (2014).

Successions of late Paleocene-earliest Eocene age comprise siliciclastic and carbonate strata that interfinger laterally and alternate vertically, and are interpreted to be of shallow-marine and continental origin. A suite of continental and paralic clastic deposits that takes the informal name of “Garumnian” is attributed to the Tremp Group, which is composed of the Thanetian to Ilerdian age Esplugafreda and Claret formations (Cuevas 1992; Rosell et al. 2001; Pujalte & Schmitz 2005; figure 2). The Esplugafreda Formation is 165-350 m thick, and largely consists of red mudstones with intercalated arenaceous to conglomeratic bodies and evaporites of local occurrence, which are interpreted to have accumulated in continental settings (Cuevas 1992; Dreyer 1993; Rosell et al. 2001; Pujalte & Schmitz 2005; Pujalte et al. 2014, and references therein). A prominent erosive surface, related to the incision of a network of lowstand valleys, defines the boundary between the

Esplugafreda and Claret formations (Pujalte et al. 2014). The Claret Formation is 10 to 70 m thick, and is composed of a varied suite of mudstones, sandstones, and conglomerates, with local gypsum accumulations, which are interpreted to have deposited in terrestrial to paralic settings (Pujalte & Schmitz 2005; Pujalte et al. 2014). Pujalte et al. (2014) describe five informal members within the eastern domain of the Claret Formation (cf. Baceta et al. 2011): member 1 consists of the infill of the basal valleys; member 2 consists of a dominantly conglomeratic unit (Claret Conglomerate); member 3 consists of yellowish pedogenized silty mudstones with intercalated sandstone bodies; member 4 is a gypsiferous unit; member 5 is mostly composed of red mudstone. The top of the Claret Formation is marked by a conformable contact with transgressive marine limestones above. Overall, the terrestrial deposits of the Garumnian strata record sedimentation in alluvial environments bordering ephemeral lakes and shallow seas; sedimentary bodies identified in the succession are interpreted as preserved geomorphic features such as channels, mid-channel bars, levees, crevasse splays, terminal frontal splays (Dreyer 1993; Schmitz & Pujalte 2007; Pujalte et al. 2014). Some of the sand-prone and conglomeratic bodies in this succession are interpreted by Mutti et al. (1996; 2000) as the preserved product of flood-dominated deposition at the termini of alluvial channels in shallow lakes. Garumnian outcrops indicate southward to westward alluvial drainage, from the structural reliefs in the north-east of the basin (Cuevas 1992; Vergés & Burbank 1996; Rosell et al. 2001; Pujalte et al. 2014). During the Paleocene, the Garumnian streams had their major source of sediment in catchments draining Variscan and Cadomian plutons of the Axial Zone of both the central and eastern Pyrenees, which were being exhumed at a rate of ca. 0.5 km/Myr (Whitchurch et al. 2011; Filleaudeau et al. 2012). These northern catchments were partly hosted in the advancing upland thrust sheets (Rosell et al. 2001). Carbonate clasts of lower and upper Cretaceous affinity are abundant in north-easterly sourced Garumnian sandstones and conglomerates, and are related to source rocks uplifted 10 to 20 km north of present-day outcrops (Schmitz & Pujalte 2007; cf. Teixel & Muñoz 2000; Gómez-Gras et al. 2016). The Montsec thrust sheet to the south represented only a minor source of sediment for the Tremp-Graus Basin in the early Eocene (Williams & Fischer 1984; Williams 1985). The clay-mineral composition of Garumnian palaeosols (Schmitz & Pujalte 2003) and of correlative deep-marine deposits (Schmitz et al. 2001) is characterized by an increase in kaolinite content recorded at the PETM, which is interpreted as related to enhanced rates of upland erosion that mobilized buried kaolinite-rich Mesozoic sediments (Schmitz et al. 2001).

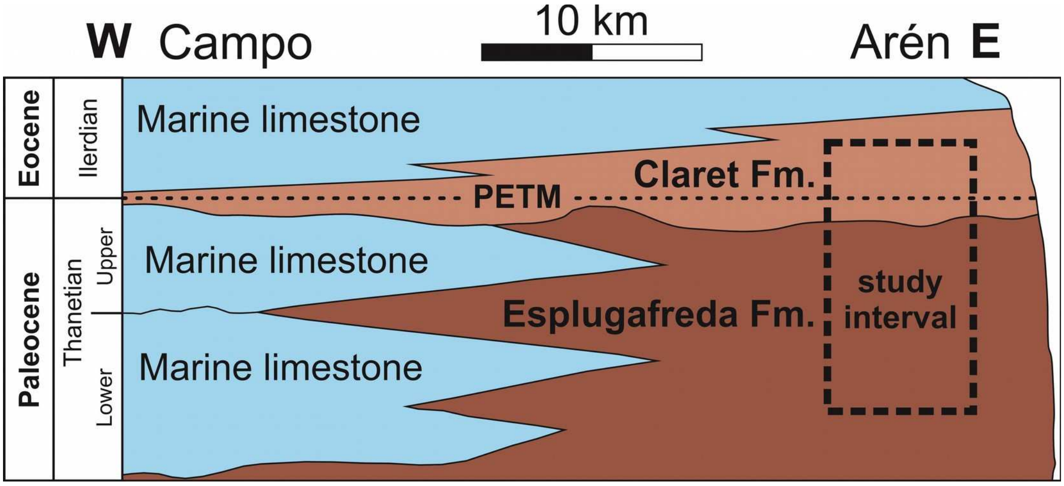


Figure 2: simplified stratigraphic scheme of the early Paleogene Garumnian along an east-west transect of the Tremp-Graus Basin, between Campo and Arén. The stratigraphic coverage of the dataset is reported. Modified after Pujalte et al. (2014).

In the Paleocene, the Tremp-Graus Basin, which connected westward to the sea, was subject to relative sea-level fluctuations of various orders (Rasser et al. 2005), the stratigraphic record of which is in part reflected in the sequence stratigraphic organization of the succession, as recorded in the Campo, Navarri and Serraduy sequences of Luterbacher et al. (1991) (cf. Eichenseer & Luterbacher 1992; Payros et al. 2000; Minelli et al. 2013). In the area where Garumnian deposits crop out, two main marine transgressions occurred in the early and late Thanetian. A widespread regression occurred in the latest Thanetian is expressed as the boundary between the Esplugafreda and Claret formations. This was followed by a major transgression during the early

Ilerdian (Rasser et al. 2005; Pujalte et al. 2014). In the Garumian domain, one of the pre-PETM valley fills of the member 1 of Pujalte et al. (2014) contains co-occurring charophyte oogonia, small benthic foraminifera, rare gastropods and *Ophiomorpha* burrows, which collectively indicate fresh to brackish salinity (Pujalte et al. 2014). The PETM, which was associated with a period of eustatic rise (Miller et al. 2005; Sluijs et al. 2008), coincided with an interval of the early Ilerdian relative sea-level rise (Pujalte et al. 2014). However, despite the development of an overall early Ilerdian transgressive phase, conditions of depositional regression are observed for the PETM interval, supposedly in relation to a climate-driven increase in clastic supply (Minelli et al. 2013; Pujalte et al. 2014).

At the PETM, the Tremp-Graus Basin was situated at a palaeolatitude of ca. 35°N (Butterlin et al. 1993), which corresponds with the position of present-day subtropical high-pressure cells. Pedogenized mudstones of the pre-PETM Esplugafreda Formation are dominantly red and purple, display colour banding and mottling, contain well-developed 0.5 to 2 m-thick Bk horizons with abundant cm-sized calcite nodules, are typically rich in Microcodium, and locally show vertic features; several gypsiferous intervals also occur in the upper half of the Esplugafreda Formation (Eichenseer & Luterbacher 1992; Dreyer 1993; Schmitz & Pujalte 2003; 2007). Overall, these observations are interpreted as indicative of a well-drained substrate in a semiarid to arid palaeoenvironment (Eichenseer & Luterbacher 1992; Schmitz & Pujalte 2007), which is thought to have been associated with dominantly ephemeral discharge regimes (Dreyer 1993). Based on independent evidence provided by clay mineral associations, Garumnian palaeosols have been interpreted as indicative of a hot climate with seasonal rainfall for the early Thanetian (Arostegi et al. 2011). The PETM interval in the Claret Formation is characterized by palaeosols with colour varying from reddish-brown, grey-purple to yellowish, and by high calcite content and dispersed carbonate nodules, indicative of a cumulate soil profile (*sensu* Wright & Marriott 1996; Baceta et al. 2011). A semiarid climate and variations from well to moderately or poorly drained conditions are inferred on the basis of these characteristics (Schmitz & Pujalte 2007; Dawson et al. 2014). The development of poorly-drained, reducing conditions in relation to a general phreatic rise is in accord with the fact that portions of the Claret Conglomerate are locally abundant in coaly fragments, found in grey mudstones and silty sandstones in the Arén area (cf. Dreyer 1993; Schmitz & Pujalte 2003). However, carbonaceous material and grey palaeosol colours are also observed in parts of the member 1 of the Claret Formation predating the CIE (Dawson et al. 2014; original field observations), suggesting that if these characteristics recorded an overall wetting trend (cf. Schmitz & Pujalte 2003), this would have been established before the PETM. An increase in precipitation amounts and seasonality at the PETM has been postulated for the broader region on the basis of numerical simulations, which point to enhanced moisture transport from the Tethys, particularly in summer months (Winguth et al. 2010). Gypsum accumulations of the member 4 of the Claret Formation, supposedly contemporaneous with the PETM recovery phase (Domingo et al. 2009; Pujalte et al. 2014), have been related to deposition in saline lakes and surrounding mud flats, indicating that a relatively arid climate was established at that stage (García Veigas 1997; Pujalte et al. 2014).

Tropical forests dominated the Iberian region during the late Paleocene-early Eocene (Barrón et al. 2010). In the Garumnian strata, rhizoliths are observed throughout the succession in muddy palaeosols, including the yellowish palaeosol of PETM age (Pujalte et al. 2014). Correlative deep-marine deposits in the down-dip Basque Basin display a turnover in the composition of the non-marine palynomorph assemblage at the PETM, suggesting that in the nearby continental areas pre-PETM permanent gymnosperm forests were replaced by a seasonal cover of angiosperms and ferns and allies, which grew during rainy episodes (Schmitz et al. 2001).

A chronological framework for the Garumnian is provided by data on the distribution of charophytes (Feist & Colombo 1983), palynomorphs (Médus & Colombo 1991) and mammal teeth (Lopéz-Martínez & Peláez-Campomanes 1999). For the Thanetian, a further chronological constraint is provided by tentative stratigraphic correlations with marine strata (e.g., Luterbacher et al. 1991). The identification of the PETM is aided by recognition of the characteristic negative carbon isotope excursion (CIE), which is recorded in palaeosol carbonates and organic matter contained in the Claret Formation. In the Arén area, a magnetostratigraphic section spanning the CIE records only reverse polarities (Lopéz-Martínez et al. 2006). Although discrepancies exist regarding the precise position of the PETM onset relative to the Claret Conglomerate (Schmitz & Pujalte 2007; Domingo et al. 2009; Manners et al. 2013; 2014; Adatte et al. 2014; Pujalte et al. 2014), the CIE can be identified in the members 2 and 3 of Pujalte et al. (2014). Although the lowest values of $\delta^{13}C$ are recognized within the member 3 by Adatte et al. (2014), these authors interpret this particular $\delta^{13}C$ excursion as the preserved signal of the subsequent Eocene Thermal Maximum 2, and argue that the position of the record of the onset of the PETM is ca. 15 m below the top of the Claret Conglomerate. However, this interpretation contradicts existing stratigraphic constraints (cf. Pujalte et al. 2009, and references therein).

Methods and data

Sedimentological data have been acquired from a ca. 7-km-long, near-strike-oriented transect from the Arén area (figure 1). Facies analysis and architectural-element analysis of fluvial deposits have been undertaken, with particular attention to fluvial channel deposits (cf. Dreyer 1993). For measured vertical sections and architectural panels, lithofacies have been described and classified in terms of textural and structural properties; Munsell colours were recorded for pedogenically modified mudstones. Palaeocurrent directions (N = 536) have been determined for individual lithofacies from cross bedding, cross lamination, clast imbrication and sole marks. The external geometry of some sedimentary units (depositional and architectural elements; see below) has been quantified through collection of data on maximum thickness and width of each unit. The values of width of the bodies are apparent; whereas the orientation of the outcrop transect is relatively constant and approximately directed along strike; stratigraphic changes might still reflect temporal or local variations in drainage direction. The internal stratal geometries of sand-prone and conglomeratic lithosomes, and the nature and hierarchical arrangement of associated bounding surfaces have also been described. The spatial relationships between sedimentary units have been recorded. Data on maximum extraclast size (N = 91) and cross-set thickness (N = 132) have been recorded for several architectural elements. The sedimentological data have been collated into a relational database that permits the digitization of the sedimentary architecture of fluvial depositional systems (the Fluvial Architecture Knowledge Transfer System, FAKTS; Colombera et al. 2012; 2013). The classification schemes employed in FAKTS have been adopted, which apply to sedimentary units that belong to three scales of observation and are termed ‘depositional elements’, ‘architectural elements’ and ‘facies units’, in order of descending scale. On the basis of their facies associations and architectural characteristics, the sedimentary bodies of the Tremp Group have been classified according to interpretative depositional- and architectural-element types. Depositional elements are large-scale sedimentary bodies classified as ‘channel-complex’ or ‘floodplain’ elements on the basis of the interpreted origin of their deposits, and the subdivision of the stratigraphy in these units is partly based on geometrical rules (cf. Colombera et al. 2012; 2013). A channel complex is a discrete channelized body, and does not possess a particular genetic or palaeo-geomorphic significance: for example, a channel complex could correspond to the preserved product of a channel belt, of a channel, to a portion of valley fill, or to a compound amalgamated multi-storey body.

Architectural elements represent sedimentary bodies with characteristic facies associations and architectural properties that make them interpretable as types of sub-environments, which generally have a distinctive geomorphic significance and commonly record the morphodynamic evolution of a particular landform. The attribution of a particular element type follows criteria based on the characters of their bounding surfaces, their geometry, scale, internal organization, and – in some instances – their relationships with other elements (cf. Miall 1996; Colombera et al. 2013). Architectural-element types include aggradational channel fills (either active or abandoned at the time of filling) and barforms (classified on dominant direction of accretion: lateral, downstream). Facies units represent packages with sub-bed-scale resolution characterized by given textural and structural properties on which they are classified. Facies units are delimited by bounding surfaces that mark a change in lithofacies, a major change in palaeocurrent, or erosional contacts (cf. 2nd-order surfaces of Miall, 1996; see Colombera et al. 2013). The adopted classification scheme of facies-unit types extends the scheme of Miall (1996). Facies types recognized in channel deposits of the study succession are summarized in table 1; photographic examples of the main lithofacies of channel deposits are presented in figure 3.

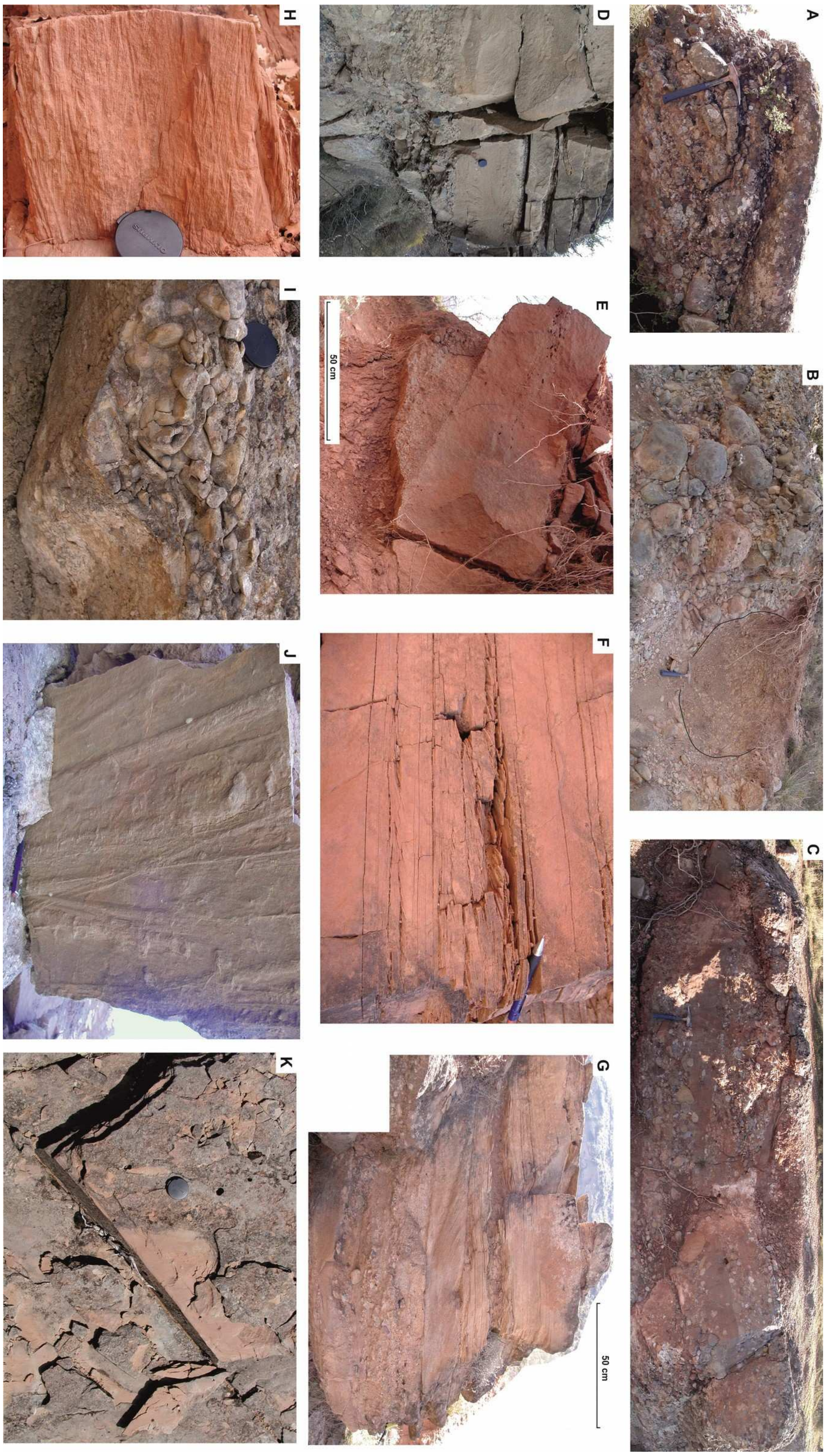


Figure 3: selected field photographs of lithofacies forming channel deposits in the Esplugafreda and Claret formations. (A) Planar cross-stratified and crudely horizontally bedded conglomerates (syn-PETM Claret

Conglomerate). (B) Clast-supported, massive boulder conglomerate; a boulder-sized intraclast that exceeds 1 m in diameter is outlined in black (deposit from member 2 or uppermost member 1 of the Claret Formation based on attribution of Pujalte et al., 2014). (C) Interbedded crudely horizontally bedded and cross-stratified conglomerates and low-angle cross-stratified sandstones (Claret Conglomerate). (D) Crudely bedded cobble conglomerates and massive sandstones (pre-PETM member 1 of the Claret Formation). (E) Crudely bedded pebble conglomerates and massive sandstones (pre-PETM Esplugafreda Formation). (F) Interbedded planar horizontally bedded and ripple cross-laminated sandstones (Esplugafreda Formation). (G) Facies forming an aggradational ribbon channel fill of the Esplugafreda Formation, comprising beds of interbedded horizontally crudely bedded conglomerates and massive sandstones, planar, trough and low-angle cross-stratified sandstones, and planar horizontally bedded sandstones; view oriented along mean palaeoflow direction. (H) Climbing ripple cross-laminated sandstone bed (Esplugafreda Formation). (I) Gutter cast from the base of a channelized unit (Claret Conglomerate). (J) Groove casts from the base of a sandstone bed (Esplugafreda Formation). (K) Plan view of a planar horizontally bedded sandstone with primary current lineation (Esplugafreda Formation). The hammer is 35 cm long; the lens cap is 5 cm in diameter; the pen is 14 cm long.

Table 1: facies types recognized in channel deposits of the study succession, based on the scheme of Miall (1996). These facies are adopted in this work to describe channel deposits only, in the studied succession.

Code	**Characteristics**
Gcm	Clast-supported, massive conglomerate
Gh	Clast-supported, horizontally- or crudely-bedded conglomerate; possibly imbricated
Gt	Trough cross-stratified conglomerate
Gp	Planar cross-stratified conglomerate
St	Trough cross-stratified sandstone
Sp	Planar cross-stratified sandstone
Sr	Current ripple cross-laminated sandstone
Sh	Horizontally bedded sandstone
Sl	Low-angle (<15˚) cross-bedded sandstone
Sm	Massive sandstone; possibly locally graded or faintly laminated
Fl	Interlaminated very-fine sandstone, siltstone and mudstone, locally with thin cross-laminated sandstone lenses

Fsm	Massive or laminated siltstone and mudstone

Sedimentological change at the PETM is analysed quantitatively. Features that are compared include the geometry and proportion of channel complexes, architectural elements and facies units. Channel complexes are not related to any given geomorphic form by definition; thus, interpretation of changes in channel-complex properties in terms of geomorphic change can only be made in consideration of changes observed at the scale of the architectural elements. Estimation of the bankfull depth of formative channels is commonly attempted from measurement of the geometrical properties of preserved deposits (cf. Bridge & Tye 2000; Mohrig et al. 2000; Leclair & Bridge 2001; Bhattacharya & Tye 2004; Hajek & Heller 2012; Lunt et al. 2013). The thickness of barforms and channel fills provides a proxy for the maximum bankfull depth of their formative channels (cf. Bridge & Tye 2000; Mohrig et al. 2000; Bhattacharya & Tye 2004). However, such estimates can be affected by issues of partial preservation, related to erosional truncation of portions of sedimentary units, and compaction. Furthermore, the hydraulic geometry of a channel as inferred from preserved deposits is not necessarily indicative of the water discharge associated with a river system, as a multi-thread river pattern (braided, anastomosing), or local changes in drainage pattern (distributary as opposed to tributary) may have developed. Additional uncertainty is associated with the correct interpretation of architectural units as preserved geomorphic elements. For example, overestimates of channel depth could arise from the misinterpretation of scour fills (e.g., confluence scours) as channel fills (Miall & Jones 2003), whereas underestimates of bankfull depth could arise from the misinterpretation of the upper portions of barforms as overbank deposits (Latrubesse 2015).

Two-sample t-tests were performed to assess whether the differences between architectural parameters (log-transformed to meet the requirements of normality and homoscedasticity) of syn-PETM and pre-PETM deposits were statistically significant. Variations in the relative proportions and grainsize of lithofacies types are employed to infer variations in the relative dominance of different depositional processes, such as the importance of bed-load, suspended-load or mass-flow deposition, or of upper versus lower flow-regime conditions. All results carry the fundamental uncertainty that is inherent in the methods of facies and architectural analysis, as applied to the rock record to infer past depositional and geomorphic processes.

Data on syn-PETM strata of the members 2 and 3 (Pujalte et al. 2014) of the Claret Formation are separately compared with corresponding data from the pre-PETM member 1 and Esplugafreda Formation.

Overall, the dataset comprises of data on:

- 247 channel complexes;
- 123 interpreted architectural elements, of which 108 are barforms and channel-fills;
- 1,397 facies units contained in channel complexes.

Results

Variations in sedimentary architecture are evaluated quantitatively across the pre-PETM and syn-PETM intervals of the Claret Formation, and also relative to the older Esplugafreda Formation as a control of the importance of stratigraphic changes through the record of the PETM.

The pre-PETM interval of the Claret Formation (member 1) is recognized as the infill of lowstand valleys (Baceta et al. 2011; Pujalte et al. 2014). Three discrete depression fills are recognized in the study area by Pujalte et al. (2014), one of which, the easternmost, appears as the compound infill of multiple coalescing depressions. The existence of the single westernmost depression, at least in the extent mapped by Pujalte et al. (2014), is here disputed based on field evidence consisting of surface correlations walked out on outcrop and traced on panoramic outcrop photomosaics, aided by recognition of channel-body pinch-out positions and palaeosol intervals. On this basis, it can be suggested that at least part of the stratigraphic interval indicated as 'western depression' by Pujalte et al. (2014) may constitute the preserved expression of an interfluve – rather than a valley – rich in channel deposits that are more ancient than the member 1 of the Claret Formation. This has implications concerning the stratigraphic attribution of deposits contained in this interval, which would predate the time of incision of the valleys at the base of the Claret Formation. However, in view of the current uncertainty and to avoid confusion with the existing stratigraphic scheme, these deposits are assigned to the member 1 of the Claret Formation following the usage of Pujalte et al. (2014).

In each of the three intervals considered, fluvial channel complexes are recognized. Each channel complex represents a discrete channelized unit, made of channel deposits (figure 4). The studied channel complexes are interpreted to relate to a range of formative processes and have variable geomorphic significance. For example, in the Esplugafreda Formation channel bodies are locally seen to form the complete aggradational infill of depressions, whereas other occurrences might just represent simple isolated channel fills, channel belts or amalgamated channel belts, locally positioned on valley floors in the member 1 of the Claret Formation.

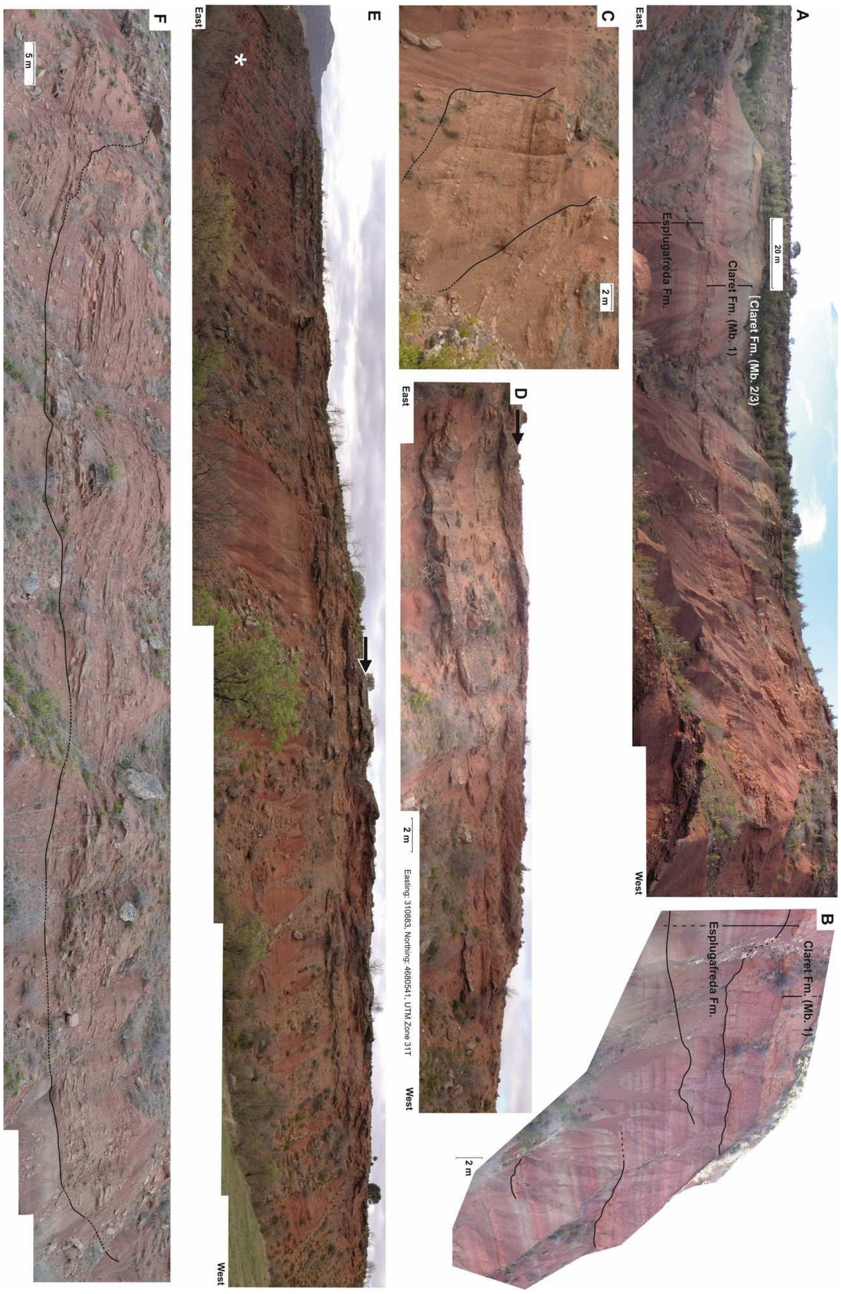


Figure 4: selected field photographs documenting aspects of the large-scale sedimentary architecture of the Esplugafreda and Claret formations. (A) Photopanel with view of the stratigraphy for the study interval on the southern hillside of the Esplugafreda valley. The member 1 of the Claret Formation in this area forms a palaeovalley fill, the 'central depression' of Pujalte et al. (2014). The stratigraphy dips into the hillside. (B) Detailed view of the sedimentary architecture seen on the slope corresponding to the western side of the photopanel in part A. The bases of cut-and-fill units from the Esplugafreda Formation are outlined in black.

The view is rotated to the approximate palaeo-horizontal. (C) Vertically stacked sand-prone channel bodies; their left-hand margins are outlined in black. (D) Conglomeratic bodies interbedded with red mudstones, seen across the transition between inferred pre- and syn-PETM deposits. The topmost conglomeratic unit corresponds to the syn-PETM Claret Conglomerate (black arrow). Some of the underlying cut-and-fill conglomeratic units are stratigraphically positioned in the pre-PETM member 1 of the Claret Formation according to the scheme of Pujalte et al. (2014), supposedly next to the western margin of the 'west depression' of the Arén area. In this area, surfaces that could unambiguously be identified as palaeovalley margins were not observed, and the variety of palaeosol colours seen in the member 1 further to the east is lacking. (E) Photopanel with view of the stratigraphy for the study interval, on a hillside of a ravine 1 km southwest of Arén. The lower section consists of deposits of the Esplugafreda Formation; the cliff is topped by deposits of the Claret Conglomerate (black arrow); the lower ledge-forming conglomeratic bodies contained in the upper section below the Claret Conglomerate are assigned to the member 1 of the Claret Formation ('western depression') of Pujalte et al. (2014). The asterisk marks the position of the channel body in figure 4F. (F) sand-prone channel body; the channel-body margins, which are expressed as erosional cuts of pedogenically modified mudstones, and its base are outlined in black.

At a lower scale of observation, different types of architectural elements are documented in the channel complexes. These sedimentary bodies are interpreted as the product of infill of fluvial channels (figure 5) and of accretion of barforms of different types (figure 6).

Two main types of barforms are identified based on lithological and architectural characteristics. In some cases inferences of the direction of accretion of these bars are uncertain, being hampered by the limited occurrence of palaeocurrent indicators and/or by the nature of the outcrop exposures. Conglomeratic barforms that appear to be dominantly accreting downstream (figure 6A-C) are seen in all three stratigraphic intervals. Some of these deposits are interpretable as bank-attached bars because of their adjacency to preserved cut-banks. Barforms that are variably gravel-, sand-, and silt-prone and that locally demonstrate lateral accretion are recognized in the syn-PETM interval of the Claret Formation, in accordance with observations by Schmitz & Pujalte (2007). Characteristic differences in facies architecture are seen between the architectural-element types that compose the channel complexes (figure 7).

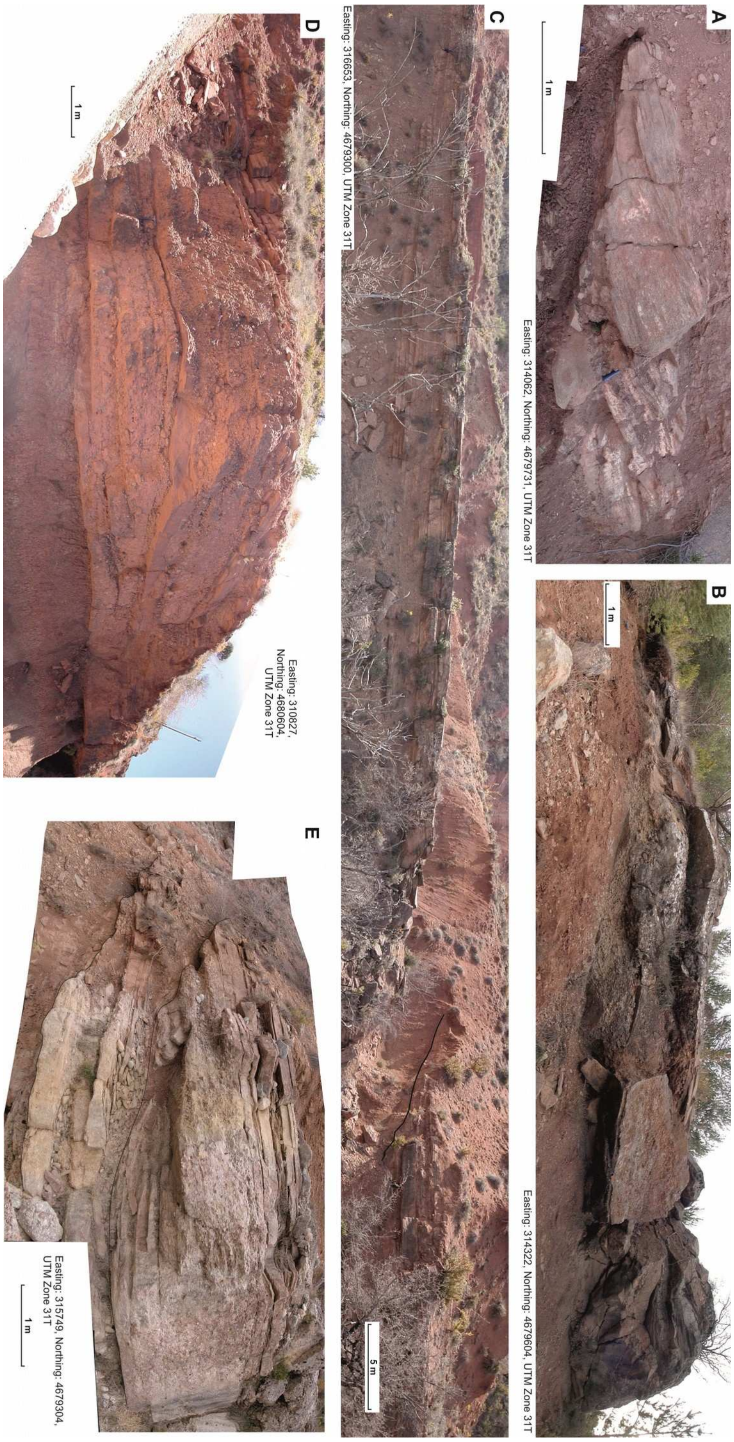


Figure 5: selected field photographs of the sedimentary architecture of aggradational channel fills. (A) Detail of the margin of a sand-prone channel-fill (CH) architectural element. (B) Detail of a channel-body margin from the Claret Conglomerate interval. The channel-body top is contained within grey mudstones of the syn-PETM member 2 of the Claret Formation; the base of the body is incised into red mudstones of the member

1. The mean palaeoflow is oriented obliquely into the outcrop. (C) Aggradational ribbon channel fill, seen to be cut both perpendicularly (to the right) and longitudinally (to the left) relative to its axis; the left-hand channel-fill margin is outlined in black. (D) Dominantly conglomeratic channel fill, with overall fining-upward trend. (E) Vertically stacked channel fills from the pre-PETM member 1 of the Claret Formation. The bases of the channel bodies are outlined in black. The mean palaeoflow is oriented approximately into the outcrop.

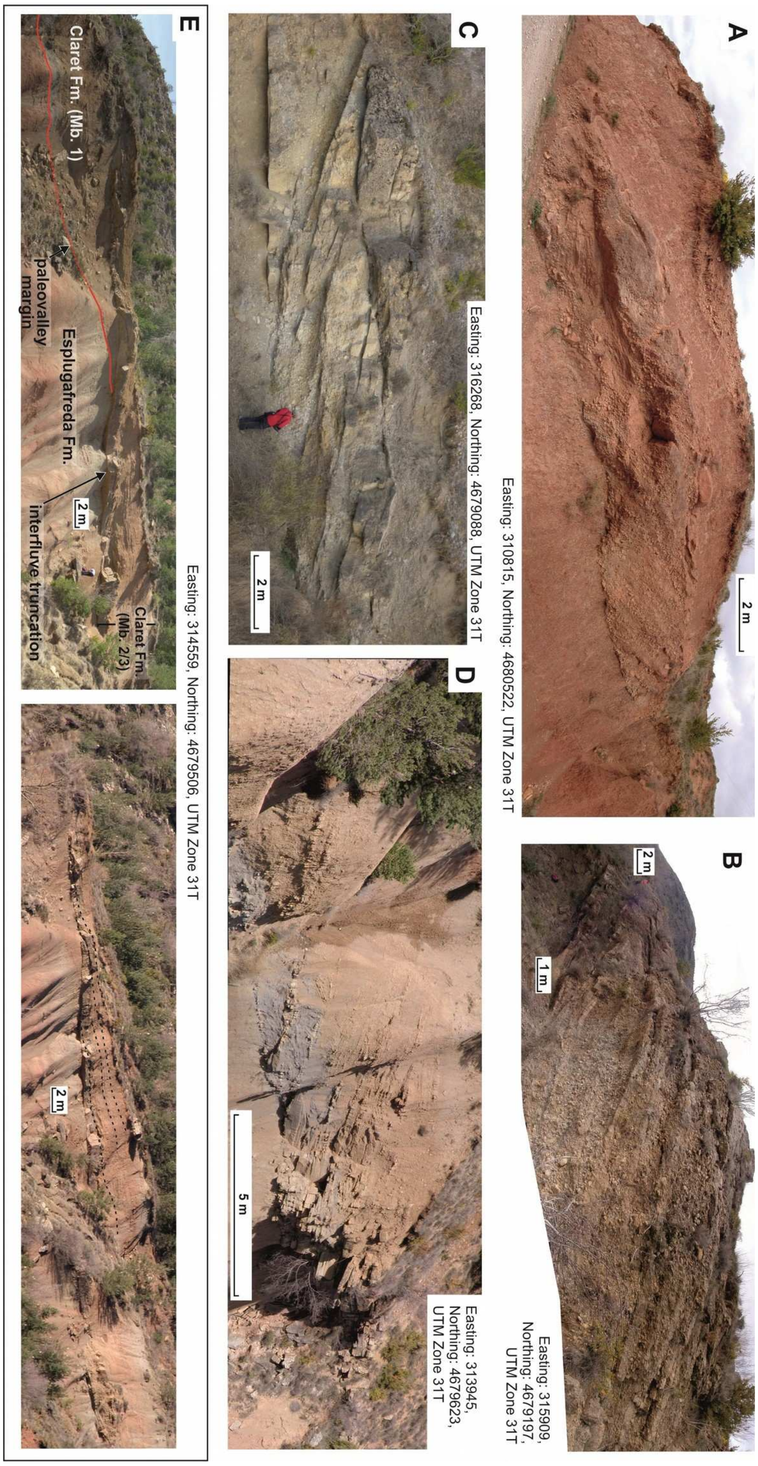


Figure 6: selected field photographs of the sedimentary architecture of barforms. (A) Gravelly barform embedded in red palaeosols from the uppermost part of the Esplugafreda Formation; see figure 2 of Mutti et al. (2000) for alternative interpretation. (B) Oblique view of a multistorey channel body from the syn-PETM

Claret Conglomerate interval. The central part of the channel body consists of a package of clinothems made of interbedded sandstones and conglomerates that is seen to record accretion at low-angle with the palaeoflow direction. The mean palaeoflow is oriented obliquely into the outcrop and to the right-hand side. (C) Multistorey channel body from the syn-PETM Claret Conglomerate interval. The central part of the channel body consists of a package with clinoforms marked by interbedded sandstones and conglomerates; this package is seen to record accretion at low-angle with the palaeoflow. The mean palaeoflow is oriented to the right-hand side. (D) Package of sandy clinothems from the syn-PETM interval of the Claret Formation (members 2 and 3). The buff-coloured upper portion of the body corresponds to the member 3 (“yellowish soils” of Pujalte et al., 2014, and references therein). (E) Outcrop expression of the relationships between the Esplugafreda Formation and the members 1 and 2 of the Claret Formation; this is the same outcrop documented in the Supplementary figure 3C of Schmitz & Pujalte (2007) and in the supplementary figure 3B of Pujalte et al. (2014). The base of the Claret Formation, consisting in the palaeovalley wall interpreted by Pujalte et al. (2014), is traced in red. Thus, the syn-PETM part of the Claret Formation is seen to rest on both the valley fill and the associated interfluve, which consists of deposits of the Esplugafreda Formation and appears as sharply truncated by deposits that mark the onset of the PETM. The ledge-forming conglomeratic unit attributed to the Claret Conglomerate may be interpreted as a basal lag; the overlying deposits consist of massive silty very fine sandstones alternating with locally pebbly, massive or faintly laminated, fine to medium sandstones; these beds are possibly genetically related to the basal lag, on which they appear to be downlapping, rather than onlapping (cf. Schmitz & Pujalte 2007 for alternative interpretations).

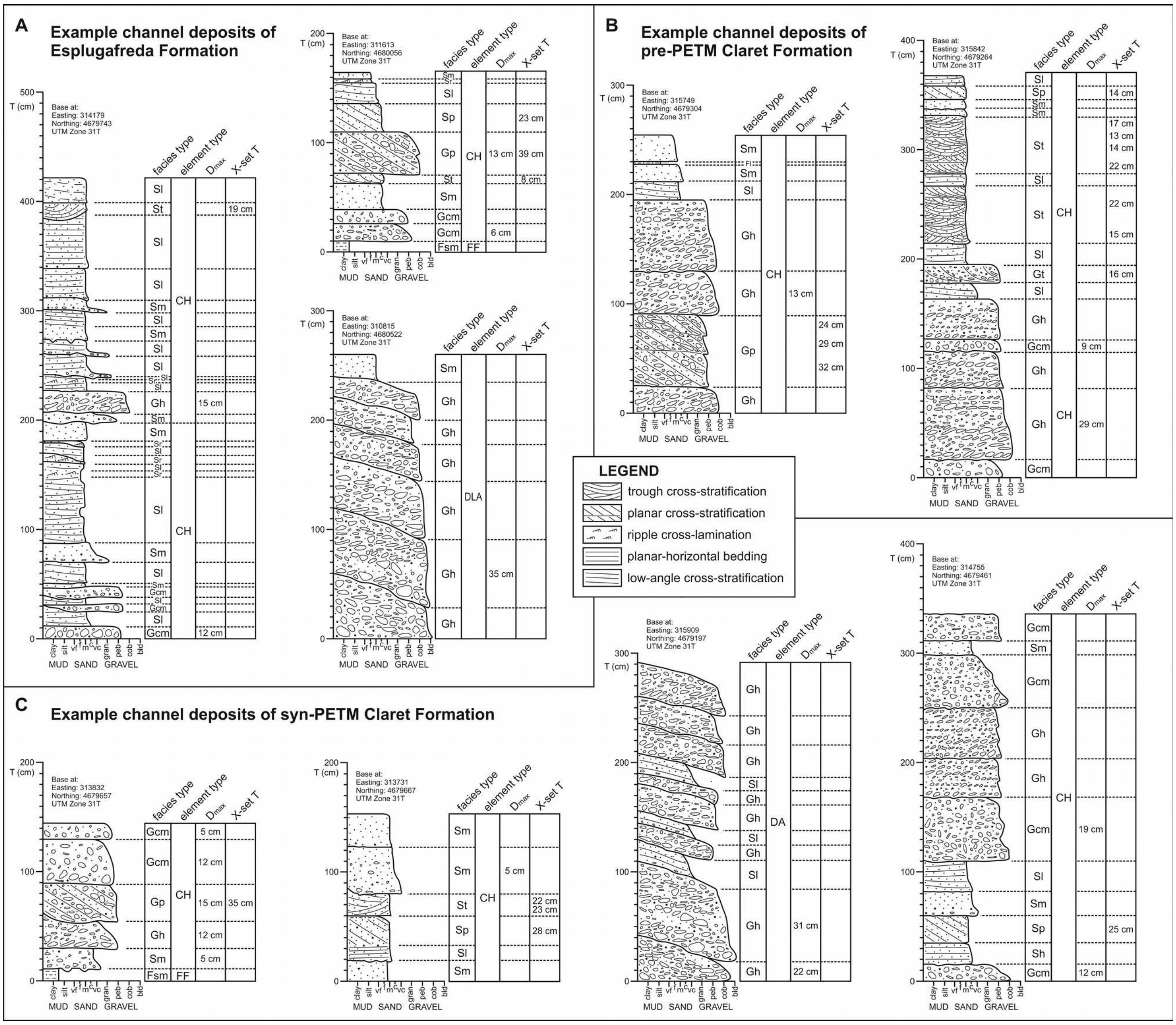


Figure 7: vertical logged sections of selected examples of in-channel architectural elements from different stratigraphic intervals: Esplugafreda Formation (A), member 1 of the Claret Formation (B), and members 2 and 3 of the Claret Formation (C). These examples have been chosen to illustrate the variability in facies organization seen in channel deposits of the studied succession of the Tremp-Graus Basin. The data contained in quantified form in the article refers to a total of 87 architectural elements, characterized through field data collection. The represented logs do not comprise all the qualitative information recorded in the field, but only what is directly made use of in this article and presented in quantified form therein. See table 1 for explanation of lithofacies codes in the 'facies type' column.

Descriptive statistics of the geometry of channel complexes from the different intervals are summarized in tables 2 and 3, and in figure 8. Channel complexes from the members 2 and 3 (N = 33) return higher values of mean, median and maximum axial thickness, compared to channel complexes from the Esplugafreda Formation (N = 186) and from the palaeovalley fills of the member 1 of the Claret Formation (N = 28). The difference in mean channel-complex thickness seen between the pre- and syn-PETM intervals is statistically significant at the 0.01 level if determined for the entire succession (two sample t-test of log-transformed data, T=-4.16, df=45, P=0.000), but only at the 0.1 level if evaluated across the members of the Claret Formation only (two sample t-test of log-transformed data, T=-1.71, df=58, P=0.093). The difference in mean channel-complex width seen between the pre- and syn-PETM intervals and determined for the entire succession is statistically significant at the 0.01 level (two sample t-test of log-transformed data, T=-4.79, df=9, P=0.001).

Table 2: descriptive statistics of channel-complex thickness for the study intervals of the Tremp-Graus Basin succession.

	Mean thickness (m)	**Median thickness (m)**	**Maximum thickness (m)**	**Thickness standard deviation (m)**	**N**
Esplugafreda Fm.	2.3	2.1	9.4	1.5	186
Mb. 1 Claret Fm.	2.7	2.7	5.0	1.1	28
Mb. 2/3 Claret Fm.	3.6	2.9	11.0	2.2	33

Table 3: descriptive statistics of channel-complex width for the pre- and syn-PETM intervals of the Tremp-Graus Basin succession. Widths are measured along a direction that approximates depositional strike, but might be apparent relative to the drainage direction of each channel complex.

	Mean width (m)	**Median width (m)**	**Maximum width (m)**	**Width standard deviation (m)**	**N**

Esplugafreda Fm. + Mb. 1 Claret Fm.	35.2	15.7	747.4	77.3	169
Mb. 2/3 Claret Fm.	484.0	352.0	1432.0	508.0	10

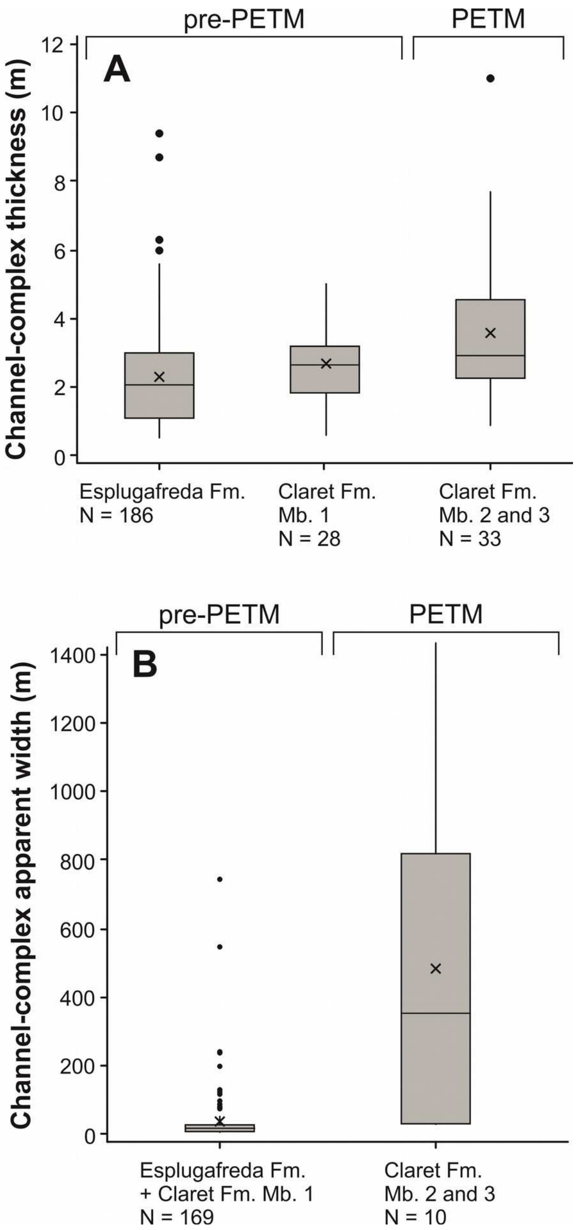


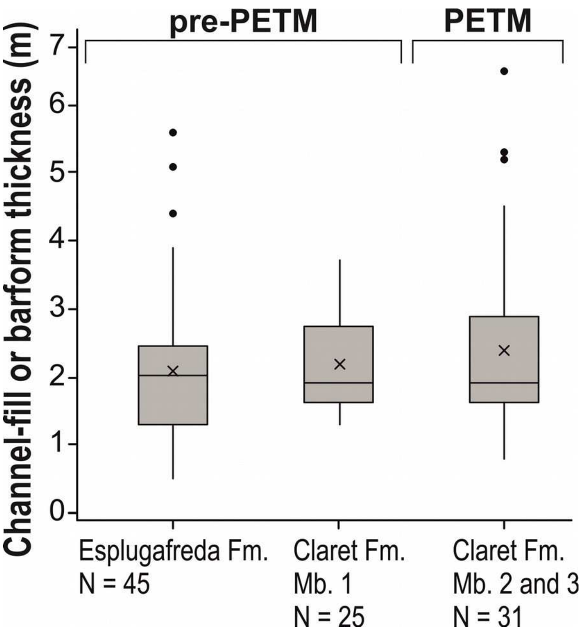
Figure 8: box plots that describe the distribution of channel-complex thickness (A) and apparent width (B) for the stratigraphic intervals considered. Boxes represent interquartile ranges, horizontal bars within them represent median values, crosses (x) represent mean values, and spots represent outliers.

Descriptive statistics of the geometry of architectural elements contained in channel complexes and interpreted as channel fills or barforms, which are present in this succession as both laterally and downstream accreting macroforms, are summarized in table 4 and figure 9. In-channel architectural elements from the syn-PETM members 2 and 3 of the Claret Formation (N = 31) return marginally higher values of mean and standard deviation of axial thickness, compared to elements from the Esplugafreda Formation (N = 45) and from the member 1 (N = 25). The difference in mean thickness for the channel fills and barforms of the pre- and syn-PETM intervals is not statistically significant (two sample t-test of log-transformed data, T=-0.82, df=50, P=0.416). A significant increase in standard deviation of channel-fill and barform thickness is seen across the pre- and syn-PETM members of the Claret Formation (Bonett's test, P=0.015; Bonett 2006). Whereas aggradational channel fills seem to dominate in the Esplugafreda Formation, over 20% of the studied in-channel architectural elements in the Claret Formation are classified as barforms: this is likely a conservative percentage, as the orientation of outcrop exposures in the Arén area commonly hinders recognition of accretion geometries.

Table 4: descriptive statistics of in-channel architectural-element (barform and channel fill) maximum thickness for the study intervals of the Tremp-Graus Basin succession.

	Mean thickness (m)	**Median thickness (m)**	**Maximum thickness (m)**	**Thickness standard deviation (m)**	**N**
Esplugafreda Fm.	2.1	2.0	5.6	1.1	45
Mb. 1 Claret Fm.	2.2	1.9	3.7	0.7	25
Mb. 2/3 Claret Fm.	2.4	1.9	6.5	1.4	31

Figure 9: box plots that describe the distribution of in-channel architectural-element thickness for different stratigraphic intervals of the Tremp-Graus Basin. Boxes represent interquartile ranges, horizontal bars within them represent median values, crosses (x) represent mean values, and spots represent outliers.



Information on the facies architecture of the channel complexes is obtained in the form of total proportions of different facies types in channel deposits, based on summed thicknesses, for the study intervals (figure 10). As compared to the Esplugafreda Formation, channel complexes from both the pre- and syn-PETM Claret Formation exhibit a larger proportion of gravelly deposits (70% versus 53%) and a smaller proportion of fine-grained deposits (2% versus 7%). In the Claret Formation, channel deposits in the syn-PETM interval display a higher proportion of conglomerates (73% versus 66%) compared to those in the pre-PETM member 1. A progressive increase (11% in the Esplugafreda Formation, to 23% in member 1, to 26% in members 2 and 3) in the proportion of cross-stratified units (Gt, Gp, St, Sp) and decrease (30% to 15% to 9%, respectively) in the proportion of plane-bedded or low-angle cross-stratified sandstones (Sh, Sl) are recorded through the three intervals. The decrease in the fraction of horizontally bedded sandstone is particularly significant across the Esplugafreda and Claret formations (28% to 4%). Through these intervals, an increase in the amount of massive sandstone in channel complexes is also recorded (3% to 8%); the proportion of massive sandstones that appear bioturbated shows modest change (12% to 14% of 'Sm' facies). Sandstones with soft-sediment deformation are notably absent from the sampled channel complexes.

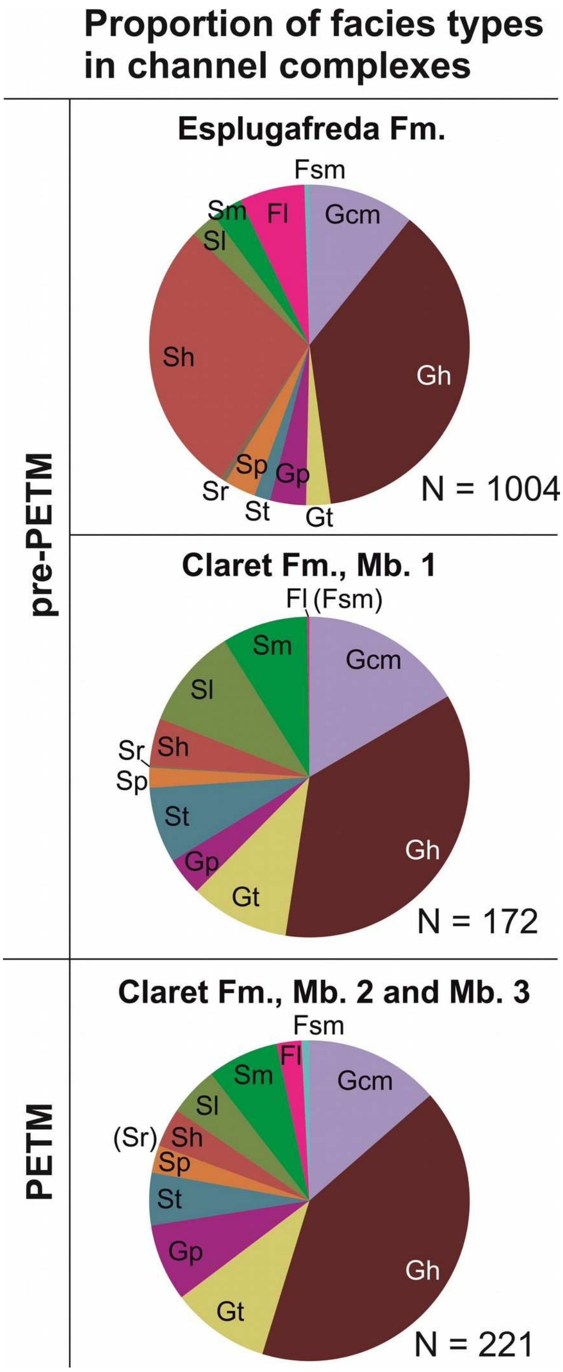


Figure 10: pie charts of the proportion of facies unit types in channel complexes from the stratigraphic intervals considered.

Descriptive statistics of measured values of maximum extraclast size by architectural element is reported in table 5 and figure 11. The largest extraclasts in architectural elements of the members

2 and 3 (N = 42) return values of central tendency and dispersion comparable with figures from the Esplugafreda Formation (N = 21) and the member 1 of the Claret Formation (N = 28). Quantitative data on intraclast size are lacking.

Table 5: descriptive statistics of maximum extra-clast size for in-channel architectural elements (channel fills, barforms) for the studied intervals.

	extraclast D_{max} mean (cm)	extraclast D_{max} median (cm)	extraclast D_{max} maximum (cm)	extraclast D_{max} standard deviation (cm)	N
Esplugafreda Fm.	22.9	23.0	45	11.4	21
Mb. 1 Claret Fm.	25.0	20.0	59	13.8	28
Mb. 2/3 Claret Fm.	23.2	20.0	63	13.6	42

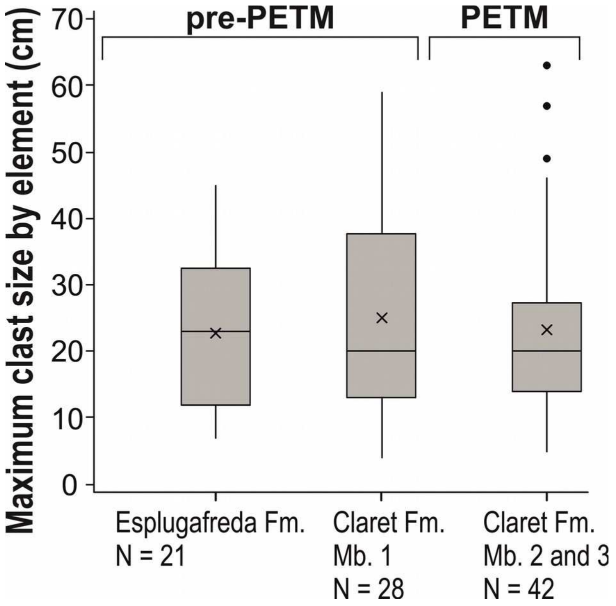


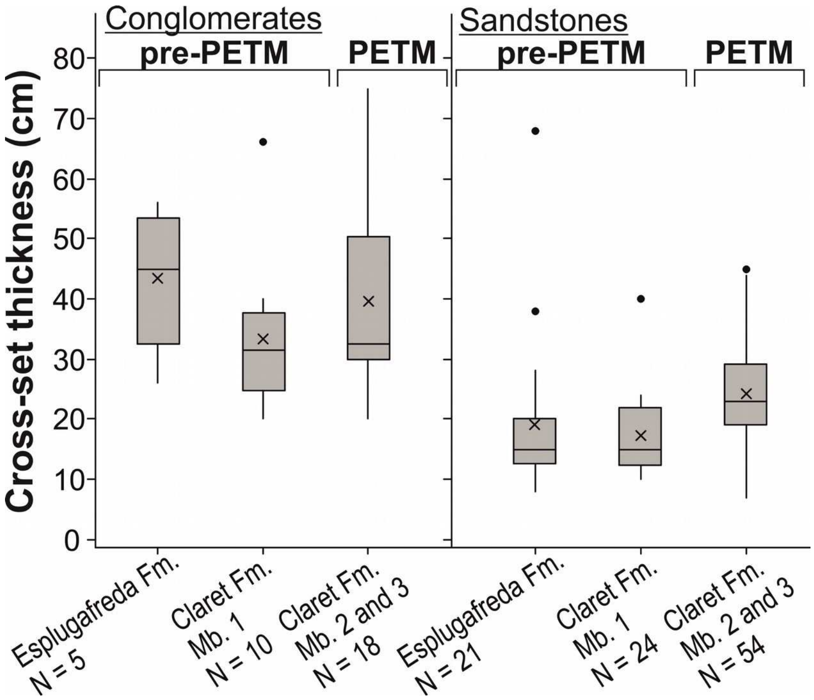
Figure 11: box plots of the distribution of maximum extra-clast size for in-channel architectural elements (channel fills, barforms) for the studied stratigraphic intervals. Boxes represent interquartile ranges, horizontal bars within them represent median values, crosses (x) represent mean values, and spots represent outliers.

Descriptive statistics of cross-set thickness have been considered for medium-scale cross-bedded conglomerates (Gp, Gt) and cross-bedded sandstones (Sp, St), and are reported in table 5 and figure 12. Mean cross-set thickness is 19.1 cm, median cross-set thickness is 15 cm, and maximum cross-set thickness is 68 cm (standard deviation = 13.1 cm). Cross-bedded sandstones from the syn-PETM members 2 and 3 of the Claret Formation (N = 54) return higher values of mean and median cross-set thickness than sandstones from the Esplugafreda Formation (N = 21) and the member 1 of the Claret Formation (N = 24). The difference in mean cross-set thickness seen in channel sandstones of the pre- and syn-PETM intervals is statistically significant (two sample t-test of log-transformed data, T=-3.51, df=96, P=0.001).

Table 6: descriptive statistics of cross-set thickness for cross-stratified conglomerates (CGL) and sandstones (SST) in channel deposits from the studied intervals.

		x-set thickness mean (cm)	**x-set thickness median (cm)**	**x-set thickness maximum (cm)**	**x-set thickness standard deviation (cm)**	**N**
CGL	**Esplugafreda Fm.**	43.4	45.0	56	11.6	5
	Mb. 1 Claret Fm.	33.9	31.5	66	12.8	10
	Mb. 2/3 Claret Fm.	39.6	32.5	75	15.6	18
SST	**Esplugafreda Fm.**	19.1	15.0	68	13.1	21
	Mb. 1 Claret Fm.	17.2	15.0	40	6.6	24
	Mb. 2/3 Claret Fm.	24.1	23.0	45	9.2	54

Figure 12: box plots of the distribution of cross-set thickness for cross-stratified sandstones and conglomerates in channel deposits from the studied stratigraphic intervals. Boxes represent interquartile



Discussion

Revisiting previous interpretations

An analysis of the significance of sedimentological change observed across the PETM in the Tremp-Graus Basin was made by Schmitz & Pujalte (2003; 2007), who also discussed the potential importance of tectonics and relative sea level. As noted by Schmitz & Pujalte (2007), the intervals that embody the onset and main phase of the PETM (i.e., the members 2 and 3 of the Claret Formation) are characterized in the Arén area by an overall higher proportion of channel deposits compared to underlying strata, and a significant reduction in channel-body density is observed vertically between the interval of members 2/3 and member 4, the latter embodying the recovery phase of the PETM (Pujalte et al. 2014).

The hypotheses that channel-body amalgamation at the PETM resulted from either subsidence reduction or relative sea-level fall can be discarded on the basis of knowledge of the inferred regime of tectonic quiescence and relative sea-level rise at the Paleocene-Eocene boundary (Pujalte et al. 2014).

Schmitz & Pujalte (2007) interpreted the increase in channel-body density seen in the member 2 (Claret Conglomerate) as representing the progradation of a braidplain that formed the proximal portion of a megafan. However, this geomorphic interpretation contrasts with converging palaeoflow directions away from the catchments (cf. Pujalte et al. 2014), and no data are available to indicate that the Claret Conglomerate represents a suite of deposits associated with a single fan, rather than with coalescing landforms. A scenario invoking a single megafan appears unlikely in consideration of the complex topography on which the member 2 accumulated, as expressed in the geometry and lateral discontinuity of the member 1. Given that these valley fills are believed to record a single phase of incision and infill (Pujalte et al. 2014), and in consideration of the limited distance between the catchments and the palaeoshoreline, the presence of multiple valley fills suggests the persistence of multiple entry points in the neighbouring mountain front during phases of widespread aggradation; the spacing of these valley fills might reflect the spacing of long-lived feeder valleys . Additionally, the deposits of the Claret Conglomerate are locally interpretable as the product of accumulation of gravelly channel lags, in relation to which overlying sand-prone deposits of the member 3 – which is largely established on pedogenic characteristics – are genetically related and synchronous (cf. figure 6E). Thus, the Claret Conglomerate alone, as a lithostratigraphic unit, cannot be interpreted in palaeogeomorphic terms.

Controls on environmental change

Intervals that encompass the PETM are also characterized by channel complexes that are, on average, slightly thicker than channel complexes from pre-PETM units. This observation likely reflects the increased degree of channel-body amalgamation, which is also expressed by a higher proportion of multi-storey and multi-lateral channel complexes in the syn-PETM member 2 of the Claret Formation, as compared to the pre-PETM member 1.

The thickness of channel fills and barforms is relatively uniform across the studied stratigraphy, which can be interpreted in terms of largely unchanged maximum bankfull depth of formative fluvial channels. If the existence of rivers with comparable size and channel-forming discharges across the PETM is assumed, the observed channel-body amalgamation could then be explained by enhanced channel mobility through faster lateral migration or more frequent avulsion (Bristow & Best 1993). However, it must be considered that the exhibited characters may also emerge in relation to the development of a network of roughly equally sized channels that form a multi-thread wandering or braided river, as opposed to a single-thread fluvial system. A braided river could accommodate a larger total water discharge and would be characterized by wider channel belts, the latter character being typically incorporated in the rock record in the form of wider channel complexes, by comparison with single-thread counterparts (cf. Schumm 1985; Gibling 2006; Colombera et al. 2013).

The hypothesis that channel-body amalgamation at the PETM resulted from increased channel mobility can be related to two fundamentally different categories of environmental change, which are not mutually exclusive:

1) changes in the drainage catchments that would drive an increase in channel mobility in the basin; such a change in river behaviour might have been caused by greater bedload delivery or reduced fine-grained suspended load delivery, which could have resulted in higher channel erosive power (cf. Nanson & Hickin 1986), faster in-channel deposition (cf. Howard 1992; Wickert et al. 2013), and perhaps decreased bank stability resulting from changes in stream-bank texture (e.g., reduced clay content; cf. Thorne 1991); increased water discharge or discharge variability could also have played a role by increasing transport flux and avulsion frequency (cf. Howard 1992; Jones & Schumm 1999).

2) changes in the nature of the depositional basin, which would permit the channels to be more mobile in relation to increased bank erodibility, for example through variations in vegetation type and density (cf. Gyssels et al. 2005), in organic-matter content (controlling soil aggregation;

Morgan 2005), or in soil drainage (positive pore water pressures reduce the effective cohesion of a soil; Thorne 1991).

Although the average thickness of barforms and channel fills shows limited change across the stratigraphy, a significant increase in thickness variability is seen across the members 1 and 2/3, which could signify more variable channel-forming water discharge during the PETM interval. Again, in view of the multi-storey and multi-lateral character of many channel complexes in the members 2 and 3, this may be related to the development of a network of variably sized channels within the braidplain setting envisaged by Schmitz & Pujalte (2007) and Dreyer (1993).

Indicative values of mean dune height and formative flow depth can be derived from cross-set thickness distributions of cross-stratified sandstones using existing empirical relationships (Allen 1970; Bridge & Tye 2000; Leclair & Bridge 2001): this approach returns estimated mean bankfull depths of 4.0 m for the Esplugafreda Formation, 3.6 m for the member 1, and 5.2 m for the members 2/3. However, the coefficient of variation of cross-set thickness suggests that the empirical relationships used are unreliable in application to the Claret Formation dataset (cf. Bridge & Tye 2000), and hence results, which would suggest increased bankfull depth during the PETM, are uncertain.

A significant change in the facies organization of channel deposits is recorded across the transition between the Esplugafreda and Claret formations. The facies associations and sedimentary characteristics of the Esplugafreda Formation channel bodies have been interpreted to be typical of a system subject to an ephemeral discharge regime (Dreyer 1993): the Esplugafreda Formation channel complexes are characterized by pause planes, a dominantly aggradational channel-fill style connected with lack of evident barform development, and a relatively high proportion of plane-bedded or low-angle cross-stratified sandstones (Sh, Sl), which can be related to transcritical to supercritical flow conditions (Fielding 2006). A shift from ephemeral to more perennial conditions may be recorded at the transition between the Esplugafreda and Claret formations (cf. Dreyer 1993), as evidenced by decreased frequency of pause planes (which appear to be absent from member 2), enhanced bar-form development, increase in the presence of structures relating to two- and three-dimensional dunes, and observation of cross-set thickness of sandy units being less variable and on average thicker – which may reflect dune height increase from the upper-stage plane bed to the dune stability field (Leclair & Bridge 2001). However, proportions of facies-unit types within the Claret Formation appear to vary little when the member 1 and the members 2 and 3 are compared: this suggests that the most significant change in channel-filling processes may have predated the onset of the PETM, and that the Garumnian system evolved relatively little in terms of in-channel depositional processes at the PETM compared to its immediate past. Based on inferences regarding the span of time embodied by the member 1 valley fills (i.e., in the order of 10^4 yr; Pujalte et al. 2014), the change in fluvial-channel facies observed across the Esplugafreda Formation and the member 1, and the concurrent change in palaeosol characteristics, might represent a response to the warming trend that is thought to have immediately preceded the carbon release recorded in the CIE (Secord et al. 2010; cf. Bowen et al. 2015).

Also, in contrast to what is stated by Schmitz & Pujalte (2007), no dramatic change is observed in the distribution of maximum extraclast size in channel fills and barforms across the Claret Formation members 2 and 3, relative to the pre-PETM member 1, suggesting that the Garumnian streams in the Arén area did not undergo any major variation in flow competence through the Paleocene-Eocene boundary. This is not entirely unexpected given that the member 1 channel complexes sit inside base-level-controlled incised-valley fills (Pujalte et al. 2014), and were therefore probably associated with a higher gradient than the member 2 and 3 channel complexes. Thus, the most evident architectural change across the PETM is in the degree of channel-body amalgamation, which can be related to both intra- and extra-basinal factors on formative-channel network characteristics (number of active channels, channel pattern) and mobility (in relation to channel and bank characteristics).

Variations in sediment supply calibre and rate may be associated with climate-driven changes in weathering mechanisms, rates and erodibility. A scenario of transient PETM wetting seems in accord with the presence of grey low-chroma deposits with coaly fragments in parts of the member 2 of the Claret Formation, transitional upwards to the yellowish palaeosol of member 3, and possibly representing the product of gleying related to waterlogged conditions connected to a water-table rise of climatic origin; however, the first occurrence of grey palaeosols in the study interval is recognized in the valley fills of the member 1. Intensified land erosion related to both seasonal extreme precipitation and sparser vegetation cover was postulated by Schmitz & Pujalte (2003; cf. Schmitz et al. 2001), in part on the basis of increased accumulation rates of terrigenous detritus in marine strata that are correlative to the members 2 and 3. It is significant that, whereas member 1 of the Claret Formation comprises deposits that onlap the palaeotopography that marks the base of member 1 itself, the same palaeotopography appears instead to be sharply truncated by the base of member 2, as evident near the palaeovalley margins and expressed in the horizontal continuity of the Claret Conglomerate (figure 6E). The relationship between the interfluves of the pre-PETM valleys and the PETM deposits, and in particular the geometry of the contact, have been previously described (Baceta et al. 2011; Pujalte et al. 2014), but its significance has been in part overlooked. The planation of the interfluves can be inferred to have occurred immediately prior to, or penecontemporaneously with, accumulation of member 2 (figure 13); this suggests either an increase in the erodibility of the interfluves, or in the erosive power of the Garumnian streams, at the onset of the PETM. The rapid erosional demolition of the interfluves would account for the increase in the rate of supply of fine-grained terrigenous sediment and progradation of clastic facies belts that are inferred to have occurred basinwide (Pujalte et al. 2015; 2016). The availability of sediment from neighbouring interfluves, together with the proximity of the drainage areas, might have resulted in a limited lag time between PETM onset and response of the river system in the Arén sector (cf. Manners et al. 2013; 2014; Pujalte & Schmitz 2014). Greater delivery of coarse-grained sediment to the Garumnian streams during the PETM may have enhanced channel erosive power and flow splitting and accelerated in-channel deposition, thereby increasing channel instability. However, it is unclear whether increased river mobility drove an increase in sediment supply by eroding interfluve areas, or rather the increased sediment yield determined faster river mobility that in turned favoured interfluve erosion. Additionally, the onset of seasonality in water discharge advocated by Schmitz & Pujalte (2003) could also be considered in relation to its control on avulsion frequency (cf. Jones & Schumm 1999; Leier et al. 2005; Plink-Björklund 2015). However, sedimentological evidence indicates that if a more peaked discharge regime characterized by high-magnitude events existed for the PETM, this left no distinctive signature in the lithofacies of channel deposits (e.g., in the form of variations in proportions of sedimentary structures that may represent the record of conditions transitional to or within the upper flow-regime, thought to be frequent under seasonal climates; cf. Fielding 2006; Fielding et al. 2009; Plink-Björklund 2015), and was likely established on a background of more perennial base discharge for the Claret Formation rivers as compared to rivers of the Esplugafreda Formation.

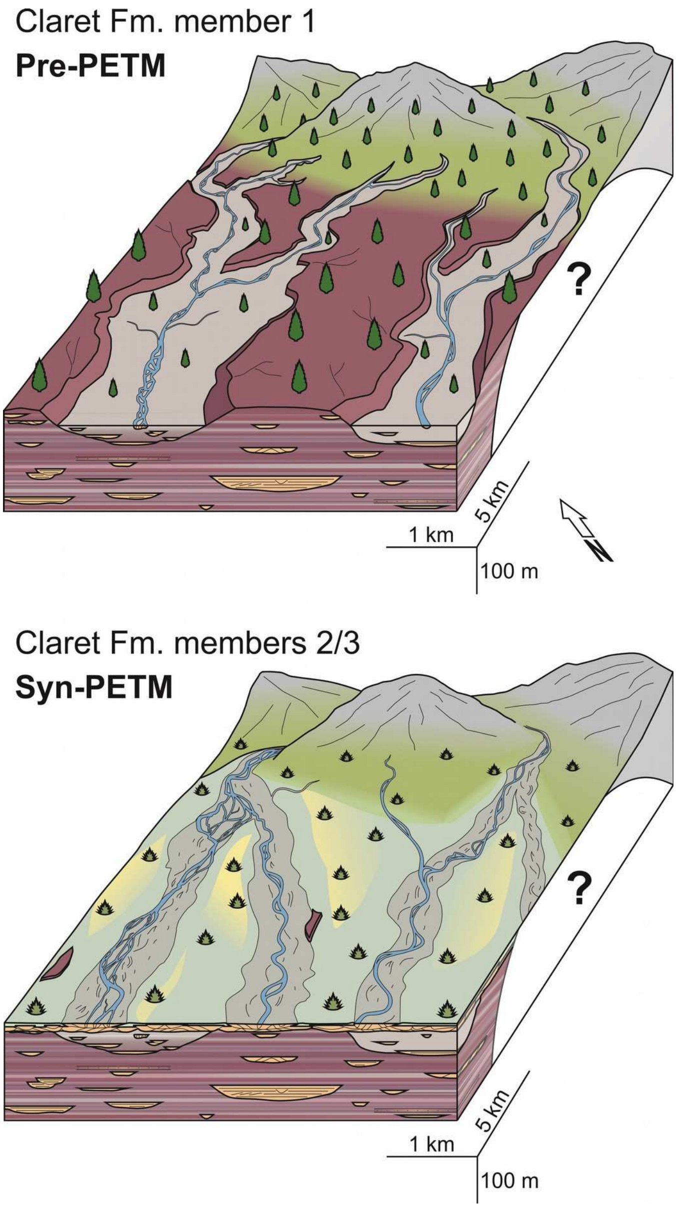


Figure 13: block diagrams that illustrate the interpreted palaeogeography for the pre- and syn-PETM intervals of the Claret Formation in the Arén area, and how the geomorphic evolution of the system across the PETM is now expressed in the stratigraphic architecture of the succession. Note the rapid transition from a stage of valley backfilling to erosional demolition of valley interfluves and widespread aggradation.

In terms of intra-basinal controls, a reduction in stream-bank stability may have resulted in response to positive pore water pressures, which reduce soil effective cohesion, or to sparser riparian vegetation. Inferences of increased channel instability and change in dominant planform morphology for the Garumnian streams at the PETM is compatible with the vegetation changeover indicated by the palynological record of the Basque Basin catchments, and consisting in the inferred replacement of permanent conifer forests with a seasonal cover (Schmitz et al. 2001).

Comparison with other fluvial systems

Other continental successions that contain a record of the PETM and of the response of mid-latitude river systems to the event are recognized in the Piceance and Bighorn basins, USA (Koch et al. 1995; Bowen et al. 2001; Burger 2012; Foreman et al. 2012). Although a detailed analysis is beyond the scope of this work, detecting commonalities between these depositional systems is useful for assessing whether variations in sedimentary architecture observed through the PETM in the Tremp-Graus Basin might reflect global or local environmental change. It is particularly significant that increased channel-body density is also seen in both the Piceance and Bighorn basins in intervals that correspond to or contain the PETM, and that lag times in the responses are similarly identified, albeit of variable duration (Foreman et al. 2012; Foreman 2014). Recognition of this particular stratigraphic signature in the different basins may reflect similar responses of fluvial landscapes and associated geomorphic processes to analogous climate-driven environmental change. Whereas a similar evolution is seen in terms of channel-body density and geometries in the Tremp-Graus, Piceance and Bighorn basins, data on facies organization (grainsize, sedimentary structures) indicate that the response of channel-filling processes to the PETM was variable in terms of both magnitude of change and type of depositional processes involved (cf. Foreman et al. 2012; Foreman 2014). This fact suggests that if a common control determined the emergence of similar large-scale stratigraphic trends, this should be a factor that dominates in controlling river mobility and size, but may be overridden by other processes in controlling depositional mechanisms. Given the current knowledge of basin histories, as constrained by available proxies on tectonic, climatic, and sea-level controls, a number of potential factors may have plausibly determined an increase in the mobility, number and/or size of fluvial channels at the PETM across all these systems: hydrological change, increased clastic influx, and variations in type and density of vegetation cover – which are themselves partially inter-related.

Conclusions

A quantitative sedimentological analysis has been carried out on outcrops of sedimentary deposits of the Tremp-Graus Basin to assess the geomorphic response of an alluvial system to environmental change through the PETM. Outcrops in the Arén area offer insight into the geomorphic change of a fluvial landscape characterized by a complex topography before the PETM. As previously recognized, the onset of the PETM marks a transition between a phase of deposition within the confines of a valley network to a phase of widespread aggradation; this transition is here interpreted to be caused by the rapid destruction of valley interfluves by syn-PETM streams, rather than by complete valley backfilling. This inference is compatible with scenarios of increased substrate erodibility or increased erosive power of the streams, at the PETM, and will likely have resulted in, and might in part have been determined by, higher rates of sediment supply. Whereas the proportion, thickness and width of fluvial channel complexes is seen to increase through the PETM stratigraphy, the thickness and maximum grain size of channel fills and barforms does not change significantly, suggesting limited variations in maximum bankfull depth and competence of fluvial channels, and indicating that the observed channel-body amalgamation might be due to higher rates of lateral migration, higher avulsion frequency, or development of a braided network. Recognition of this particular stratigraphic signature may reflect a response of the fluvial landscape and processes to types of climate-driven environmental change in accordance with what has previously been suggested for this basin, i.e., in relation to enhanced hydrological cycle, increased seasonality, and vegetation loss. However, the facies organization of syn-PETM channel deposits of the Claret Formation does not appear to be significantly different from that seen in the immediately preceding pre-PETM channel bodies of member 1.

Because the analysed characteristics of fluvial sedimentary architecture represent a record of river processes and landforms, the studied succession might represent an analogue with which to predict the geomorphic metamorphosis of river systems in certain mid-latitude regions under conditions of rapid global warming. However, further research on the Tremp Group is required to better constrain the relative importance of specific factors and to elucidate the behaviour of the river network at the scale of the entire basin.

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