Palaeoenvironments of the Late Miocene Prüedo Basin: implications for the uplift of the Central Pyrenees

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Abstract: The nature, structure and extent of a palaeo-basin sedimentary infill exposed in the Aran valley (Central Pyrenees) was studied by combining stratigraphical and biostratigraphical analyses and an audio-magnetotelluric survey. The basin developed on top of a pre-existing peneplain and was formed by the North Maladeta Fault activity. The fluvi-palustrine sequence filling the basin was at least 100 m thick. Specimens of the taxon Hippuris cf. parvica/carpé Nikitin were identified for the first time in a European palaeoflora. The palynological and carpological analyses allowed us to (1) constrain the age of the basin infill as Vallesian (11.1–8.7 Ma), (2) characterize the vegetation of the belt surrounding the basin as a mainly temperate to warm-temperate assemblage, and (3) estimate the palaeoaltitude of the site at between 700 and 1000 m, which leads to an altitude change of 900–1200 m for the North Maladeta Fault downthrown block and 1640–1640 m for the upthrown block. These data allowed us to estimate the regional uplift of the area at between 0.08 and 0.19 mm a−1. The different exhumation values obtained by other researchers for sites located at both sides of the North Maladeta Fault are in agreement with its activity as a normal fault since the Late Miocene.

Supplementary materials: Details of the data acquisition, processing and modelling for the Porèra profile are available at www.geolsoc.org.uk/SUP18551.

Fluvio-lacustrine deposits are very valuable sedimentary archives of geological processes. When located in high mountains, these deposits might be especially relevant to the reconstruction of orogenic histories (Gierlowski-Kordesch & Kelts 1994; Cohen, 2003). However, such intramontane palaeo-lakes are rarely preserved and tend to disappear over time because of the erosion dominating the surface processes in high mountains.

This paper is focused on the Prüedo Basin, a palaeo-lake preserved on the Axial Zone of the Pyrenean orogen. Since the end of the main Alpine collision, the Pyrenees have undergone a generalized period of stress relaxation with local episodes of compression. During this continuing post-orogenic period, several intramontane basins were generated along the axis of the Pyrenees. Three of them formed during pre-Quaternary times: the Arlas Graben on the Pierre-St-Martin massif (Viers 1977, as described by Fourniquet 1978; Hervouët 1997; Lacan 2008) in the western part of the range; the Prüedo Basin, located in the central part; and the Cerdanya–Conflent Basin system (Roca 1986, 1996; Agustí & Roca 1987; Cabrera et al. 1988; Martín-Closas et al. 2005; Calvet & Gunnell 2008) in the eastern part of the Pyrenees. The Arlas Graben is the result of the negative inversion of alpine thrusts, as corroborated by microtectonic data (Hervouët 1997). No record of sedimentation has been found in this small graben (Lacan 2008). In the Prüedo and the Cerdanya–Conflent Basins, part of their sedimentary infill has been preserved, representing a unique record of the post-orogenic evolution and Neogene palaeoenvironmental conditions of the chain (Fig. 1).

De Sitter (1954) has described the Prüedo deposits (Central Pyrenees) and interpreted them as the infill of a deep gorge incised in a pre-existing peneplain. In contrast, Kleinsmiede (1960) has suggested that the deposits formed before the peneplain and that they have been affected by post-orogenic east–west faults. Jelgersma (1957) has assigned a post-Vindobonian age to them (i.e. European continental Late Miocene stage) and has attributed the Prüedo Basin to surface subsidence because of the activity of the bounding faults. Alternatively, glacial ages have also been postulated for these deposits (Solé Sabarís 1957; Vilaplana et al. 1986).

Similar planar surfaces have been identified in other parts of the Pyrenees and have been classically interpreted as the remnants of a regional erosional surface generated near base level before the Quaternary fluvo-glacial incision (e.g. Birot 1937; Sitter 1952; Ortúñó 2008, and references therein). Some researchers (e.g. Babault et al. 2005, 2007) have maintained that the peneplain was generated at its current high altitude (c. 2000 m) as a result of the high elevation of its original base level (i.e. the Ebro Foreland Basin). More recently, Calvet & Gunnell (2008) have proposed that the different erosional surfaces in the Pyrenees might have different origins (from alitplanation to elevated peneplains) and have argued that the surfaces preserved in the Eastern Pyrenees have undergone >1 km uplift since their generation in Mid-Miocene times. In the Central Pyrenees, the remnants of post-orogenic peneplains have been described by Kleinsmiede (1960), Zandvliet (1960) and Ortúñó et al. (2008), but no consistent age has been assigned to them to date.
This study seeks to (1) propose a model for the generation of the Prüedo Basin and, especially, to constrain the age of the Prüedo deposits and, therefore, a minimum age for the peneplain remnants located at their base, and (2) determine the nature and palaeoenvironmental record of this Late Miocene alluvial system with the aim of establishing the palaeoaltitude during the basin formation: implications for the surface uplift of the Central Pyrenees since the Late Miocene will be discussed.

**Geological and geomorphological setting**

The Pyrenean range has resulted from the collision of the Iberian and Eurasian plates (Fig. 1). The major convergent episode started in the Late Cretaceous and lasted until the Late Oligocene in the Eastern Pyrenees, and until the Mid-Miocene in the Western Pyrenees. The structure of the Pyrenees is characterized by a thrust system displaying an asymmetric double wedge of upper crustal rocks. Both the northern and the southern wedges consist of a thrust basement and cover rocks, forming an imbricated system. The northern wedge constitutes the North Pyrenean Zone and thrusts northwards over the Aquitaine Basin, the northern Pyrenean foreland. The southern wedge thrusts southwards over the Ebro Basin, the southern Pyrenean foreland. Basement rocks in this wedge form an antiformal stack, classically known as the Axial Zone, which is composed of three main thrust sheets. The thrusts separating the basement units are folded and dip to the south in the southern limb and to the north in the northern limb of the antiformal stack. The Gavarnie thrust is one of these major thrusts. The detached cover units in the southern Central Pyrenees are known as the South Pyrenean Central Zone (Muñoz 1992; Beaumont et al. 2000; Jolivet et al. 2007; Fig. 1).

The beginning of the post-orogenic period in the Pyrenees is asynchronous, starting during the Mid-Oligocene in the eastern part and during the Mid-Miocene in the western area (Lacan & Ortuño 2012; and references therein). This period is characterized by a relatively slow convergence (e.g. Vergés 1993; Meigs & Burbank 1997; Beaumont et al. 2000; Fidalgo González 2001; Sinclair et al. 2005), with low-present-day global positioning system (GPS) rates showing uncertainties greater than the tectonic signal (\(<1\)\,mm\,a\(^{-1}\); Nocquet & Calais 2004) and a slight extensional movement continuing in the last 3 years (Asensio et al. 2012). The review of the neotectonic research published to date has led Lacan & Ortuño (2012) to classify the chain into areas recently affected by consecutive extension and compression (Eastern Pyrenees), predominant extension (Central and Western Axial Zone) and horizontal compression along a submeridian direction (Pyrenean forelands). The neotectonics of the Pyrenees, according to those workers, are the expression of the continuing uplift dominating the inner chain (High Chain) as a consequence of isostatic processes, and horizontal submeridian compression at the external chain (Low Chain) as a result of the Eurasian and African collision (e.g. Lacan et al. 2012).

The study area is located on the axis of the Central Pyrenees, between the east–west-trending Aran Valley, through which the Garona River flows, and the Aigüestortes–Maladeta Massif (3404 m maximum elevation). Basement rocks cropping out in this area consist of Palaeozoic metasedimentary rocks intruded by Late Variscan granitoids. Permo-Triassic and older rocks are faulted and preserved on top of some of the peneplain remnants, a few patches of lateral moraines and some rocky glaciers (Bordonau 1992; Serrat & Vilaplana 1992). The post-glacial sediments are limited to debris-slope deposits, some alluvial sediments in the valley bottom and a number of peat zones in the flat areas (Figs 2–4).

The North Maladeta Fault (Kleinsmiede 1960; García-Sansegundo & Poblet 1999) is one of the main structures in the region. According to Bordonau & Vilaplana (1986), it could be the normal fault displacing the glacial slopes of the Port de Vielha ridge. More recently, Ortuño (2008) and Ortuño et al. (2008) identified this structure as an inverted segment of the Gavarnie Thrust, strongly dipping to the north in the northern flank of the Axial Zone Antiformal Stack. After a structural, geomorphological and geophysical study of the area, those workers suggested that the activity of the fault led to the formation of a semigraben, the Prüedo Basin, offsetting the remnants of a pre-Pliocene peneplain by a maximum throw of \(c.\) 500 m. They identified the North Maladeta Fault as the
most probable source of the Vielha earthquake (Mw 5.8; 19 November 1923) on the basis of the macroseismicity distribution provided by Susagna et al. (1994).

Methods

We present the results of a multidisciplinary study of the Prüedo Basin sedimentary infill. The structure of the basin was constrained by the integration of geological and structural mapping and subsurface data obtained from an audio-magnetotelluric (AMT) survey (Simpson & Bahr 2005). This survey was conducted on the hanging wall of the North Maladeta Fault, with the aim of ascertaining whether there is evidence of the Prüedo deposits at depth, and to determine the geometry of its confining basin, which included the identification of the faults.

The survey was conducted along the Prüedo and Porèra north–south profiles (Fig. 3). The resistivity model obtained from the Prüedo profile has been presented by Ortuño et al. (2008). For the Porèra profile, a 2D resistivity model along a north–south line was obtained, using the RLM2DI inversion code (Rodi & Mackie 2001). The final model has a r.m.s. value of 4.8.

The stratigraphical and sedimentological characteristics of the basin infill were established by analysis of the available outcrops to determine the origin and provenance of the sediments. To establish the palaeoenvironmental conditions during the infilling of the basin, this analysis was combined with palaeontological research on selected strata. Several layers of the upper part of the sequence were sampled and analysed to determine the fossil content. One layer was sampled and processed for palynological purposes and three layers were sampled to separate and identify the carpological content.

The material sampled for palynological analysis was processed with cold HCl (35%) and HF (70%) to remove respectively carbonates and silica. Separation of the palynomorphs from the remaining residue was carried out using ZnCl2 (density = 2 gr/cm3). Sieving was carried out with a 10 μm nylon sieve. The pollen residue together with glycerin was prepared on slides. A transmitted light microscope, using ×250 and ×1000 (oil immersion) magnifications, was employed to identify and count the palynomorphs. Spores were not considered because of their low representation. Pollen grains were identified by comparing the fossils with their present-day relatives using several pollen atlases (China, Taiwan, Africa, North America, Mediterranean Region, etc.) and the Photopal website (http://medias.obs-mip.fr/photopal). A minimum of 150 terrestrial pollen grains besides Pinus and indeterminable Pinaceae were counted in each analysed sample following the procedure proposed by Cour (1974).

The carpological content was analysed following the procedure described in detail by Martinetto (2001, 2009); the sediment was broken up using a 3% H2O2 solution. Some of the carpological remains floated to the surface and were collected using a cloth net of 1.5 mm mesh. Material exceeding 3 mm in diameter was separated from the disaggregated sediment using a 0.3 mm metallic sieve, and was observed under a binocular microscope together with the material previously collected.

The age of the Prüedo deposits was inferred by comparing the pollen and carpological content of this sequence with sequences
dated in other parts of the Pyrenees and in Siberia. The palaeoaltitude estimated for the original Prüedo Basin was obtained by comparing the pollen spectra obtained with those observed in modern analogues for the Late Miocene. This method has proved successful in palynological studies of other Neogene basins (e.g. Suc 1984; Jiménez-Moreno et al. 2005, 2008; Jiménez-Moreno & Suc 2007).

A basin on top of a pre-existing peneplain

In map view, a lensoidal basin can be observed by looking at the geometry of the North Maladeta Fault trace and the location of the sites where the deposits are exposed or inferred from the audio-magnetotelluric survey (Fig. 2).

The resistivity models obtained for the Porèra profile show a number of domains that could be identified with the main rock units exposed in the area. High resistivity values (>300Ωm) are attributed to the granitic rocks of the Maladeta and Tredòs batholiths (domains A and D, Fig. 5a). Low resistivity values (10–15Ωm) correspond to the Devonian limestones, which extend to considerable depth (domain C in Fig. 5a). In the Porèra section (Fig. 5a) the relatively low resistivity near the surface of this domain can be explained by the presence of karst cavities filled with conductive clay minerals. The subhorizontal unit that shows a low resistivity (10–80Ωm; domain B in Fig. 5a) represents the Prüedo Basin infilling and is located below the Porèra high-resistivity (150–300Ωm) till cover over the granitic rocks of the Tredòs batholith. The geometry of the described domains and the surface geology allowed us to draw two north–south cross-sections across the basin (Fig. 5).

The base of the basin infill is tilted to the south, so that the depocentre is located close to the North Maladeta Fault. Such tilting was controlled by two faults: the North Maladeta Fault, which bounds this domain to the south, and the Tredòs Fault, which represents the northern limit of the tilted area. The base of the Prüedo infill has an undulating shape and could correspond to the remnants of the eroded peneplain surface.
RECONSTRUCTION OF THE PRÜEDO BASIN

Fig. 4. General view. The Prüedo peneplain viewed from the peneplain to the west. The top of the lateral moraine in the Rencules valley, the faceted spur of the North Maladeta Fault below Salana’s peak and site 1 are indicated.

Fig. 5. Porèra section. (a) Resistivity model, and geological interpretation. Resistivity contrasts observed in the model are marked by lines. White curved lines correspond to lithological changes interpreted as non-tectonic (erosive or intrusive) contacts. Bold straight white lines are tectonic contacts corresponding to the North Maladeta Fault, the Tredòs Fault and secondary faults. The geological section is based on the integration of audio-magnetotelluric and field data. (b) Geological interpretation of the Prüedo section magnetotelluric survey (in Fig. 3), modified from Ortuño et al. (2008), with the same horizontal scale as the Porèra section. The Prüedo basin is interpreted in this section as having been formed on top of a pre-existing planar surface as a result of the activity of the North Maladeta Fault and secondary faults. The maximum extension of the Neogene basin is 1.2 km.
In the Prüedo and Porèra cross-sections, the maximum extent of the basin to the north is indicated by the outcrop of the remnants of the peneplain. This northern limit is located at 1.2 and 1 km from the southern limit in the Prüedo and Porèra sections, respectively. However, the area in which most of the sediments were trapped has significantly smaller dimensions: 0.5 km at Prüedo and 0.2 km at Porèra. In both sections, this depocentre zone is locally bounded by faults related to morphological scarps affecting the till uppermost surface. In the Porèra section, the southern boundary fault corresponds to the North Maladeta Fault easternmost segment (Figs 2 and 3).

The extent of the Prüedo Basin to the south is defined by the step of the North Maladeta Fault central segment between the Aiguamòg and Valarties valleys. The stepovers between the eastern and western fault segments and the central segment might have caused secondary faulting and subsidence to accommodate the differential movement. The offset of the peneplain owing to the North Maladeta Fault is 440 m in the Prüedo transect and 515 m in the Porèra transect. The comparison between the two surveyed transects shows a greater tilt of the base of the basin in the Prüedo section (6–7°) than in the Porèra section (3–4°). In both cases, this tilting is greater than that observed for longer profiles of the peneplain remnants (c. 3°) obtained through topographic analysis across the North Maladeta Fault (Ortuño 2008). This greater tilting next to the North Maladeta Fault trace is explained by the activity of the secondary conjugate faults located in the hanging wall and next to the fault trace.

The record of a fluvio-lacustrine system

The Prüedo clastic sequence was identified at five locations, all of them situated less than 8 km north of the present east–west main watershed of the Pyrenean range. Three of the outcrops identified are located on the Prüedo plains, and two of them on the Porèra plains (Fig. 3). Sites 1 and 2 are large enough to allow the general structure of the deposits to be distinguished.

The stratigraphical description of sites 1 (Riu Verd) and 2 (Riu Merder) was undertaken. At each of these sites, several outcrops were indicated by ‘N’ or ‘S’ and a number, according to their location on the northern or southern slope of the creek (Fig. 3).

Sites 1 and 2 are exposed as a result of 40 m high landslide scars. The uppermost part of the sequence is covered by a till deposit, ranging in thickness between 2 and 50 m. The most complete record of the sequence (102 m thick) is exposed at site 2. This thickness must be considered as a minimum estimate for the complete sequence as its base is not exposed and its top is bounded by an erosive contact (the base of the till).

Facies associations (FA) and changes of sedimentary environment

The beds can be grouped into three lithofacies according to the grain size of the sediments and their stratigraphic position. These lithofacies, which are summarized in the stratigraphic columns in Figures 6 and 7, are described below. Table 1 interprets the lithofacies associations in this study according to the classification proposed by Miall (1978).

FA1, river channel deposits (conglomeratic basal unit). The base of the sequence is composed of Gm (clast-supported conglomerates) and Fl (fine sands and silts) lithofacies. The conglomerate layers are compacted, locally cemented, and interbedded with
layers of clay–silt and layers of sand (medium and coarse grain size). The clasts are made up of granite, schist, greywacke and hornfels, and show imbrications that indicate a NE to NNE provenance. This unit is best exposed in outcrop S1 at site 2 (Figs 3 and 8c). The FA1 lithofacies association is interpreted as the deposit of a river channel that locally migrated and was replaced by lateral bars.

**FA2, river flood plain and channel deposits (silty and micro-conglomeratic intermediate unit).** This assemblage is composed of Gm, Fl and Ss (fine to very coarse sands) lithofacies. It is made up of alternating layers of silt, sand, micro-conglomerates and minor conglomerates and is exposed at site 2 (Fig. 8b), where a fining upwards trend is observed in the grain size. The conglomeratic levels have erosive bases and channel-like geometries, and are not completely cemented. The main differences from the FA1 (basal units) are (1) the predominance of finer material (silt and fine gravels or coarse sand), occasionally rich in organic matter, (2) the lack of well-cemented matrix, (3) the smaller size of the clasts, and (4) the presence of granite clasts and the absence of greywacke and hornfels clasts. FA2 is interpreted as the deposit of a low-energy river channel and a flood plain environment.

**FA3, river flood plain to lacustrine environment (lignitic upper unit).** The uppermost part of the sequence corresponds to lithofacies Fl and C (lignite and lignitic muds). It consists of alternating layers of grey silt and lignite, with occasional layers of medium and coarse sand. The lignite layers are up to 1 m thick and gradually become more abundant in the uppermost part of the sequence, where a rhythmic alternation, consisting of coarse sand, silt and
Table 1. Facies description and interpretation of the Prüedo deposits (following codes of Miall 1978)

| Code and description | Lithologies and provenance | Interpretation |
|----------------------|----------------------------|----------------|
| Gm: clast-supported, interbedded massive and/or crudely bedded conglomerate with horizontal stratification and clast imbrication. Heterometric and rounded grain-supported conglomerate with planar and sub-spherical clasts <10 cm diameter | (FA1) quartzite–greywacke, schists, slates, and hornfels, with minor amounts of marble and granitic clasts; provenance north to NE | Gravel sheets (a few clasts thick) moving during peak flow, and longitudinal bars associated with episodes of high water and sediment discharge forming fining-upwards clast accumulations as the bars built up towards shallower water levels |
| Fl: thin, poorly laminated fine sands and silts with a sheet-like geometry | (FA2) granite and schist, and more rarely, quartzite; provenance NNE–ENE | Progressive abandonment of the channels, vertical aggradation |
| Sh: coarse and fine sands with horizontal laminations, forming broad sheets | | Filling of erosive scours |
| C: coal; lignites and lignitic muds | | Sandy bedforms representing broad sheets of sands associated with slack water deposits such as abandoned channel fills developed in unconfined channels |
| | | Palustrine deposits |

Palaeoenvironmental record of a Late Miocene lacustrine system

Palynological analysis

The palynomorph-rich sample of this study comes from the lignite layer Prüedo 4d, located at site 2, outcrop 4S (Figs 2 and 7). We analysed 15 g of sediment. The sample revealed 33 taxa (Table 2), with the pollen spectra indicating a flora dominated by Cyperaceae and Poaceae and other herbaceous taxa such as Amaranthaceae–Chenopodiaceae, Apiaceae, Campanulaceae, Plantago, Liliaceae, Caryophyllaceae, Rumex, Convallvulus and Astereae. *Pinus* and indeterminable Pinaceae are relatively abundant and so is *Cathaya*, a mid-altitude conifer growing today in SE China (Wang, 1961). Swampy taxa such as *Taxodium* type and *Myrica*, and mesothermic riparian plants such as *Ulmus–Zelkova, Salix, Populus, Liquidambar* and *Juglans* are present in the pollen spectra. Some thermophilous elements occur and are represented by *Engelhardia*, Sapotaceae, *Mussaenda* type, *Distylium* and *Microtropis* fossil, *Quercus* *ilex–coccifera* type, *Olea*, Oleaceae and *Quercus* deciduous type were also identified, the last one being relatively abundant. Besides *Cathaya*, the micro- and meso-microthermic elements such as conifers suggesting a high altitude (*Cedrus, Picea* or *Abies*) are very rare and were represented only by *Cedrus*.

Carpology

Three layers (indicated in Figs 6 and 7) were sampled for the carpological analysis. The layers correspond to fine sand within well-compacted layers. Approximately 1 kg of sediment was processed.

The Prüedo deposits provided anatomically preserved carpological remains. The material recovered consists of a few tens of millimetre-sized fruits and seeds from samples 1a, 4a and 4d. The fact that these fruits and seeds are mostly deformed and partly abraded together with few comparable Pyrenean Neogene carpological floras published does not facilitate identification. Nevertheless, the morphology of these specimens was enough to permit identification at genus level by comparison with analogous material described in the literature or from the palaeo-carpological collection of the Earth Sciences Department of Torino. The list of the identified carpological specimens is given in Table 3. Remains of the genera *Alisma, Carex, Ranunculus* (herbs) and *Rubus* (frutex) are common in West European Neogene and Quaternary carpological assemblages. *Weigela* is a shrub whose seeds are found in the Miocene and Pliocene of Western Europe (Mai & Walther 1988). One type of carpological remains is interpreted as the endocarp of an extinct form of the aquatic herb *Hippuris* (i.e. *Hippuris* cf. *parvicarpa* Nikitin; Fig. 9). Some of the endocarps of these specimens are almost cylindrical, c. 1.0 mm × 0.6 mm in size, with a circular hilum, centred on the apical face. Such a circular hilum, perpendicular to the long fruit axis, is characteristic of *Hippuris* rather than *Myriophillum*, which is distinguished by an oblique hilum plane (Fig. 10). Fossil endocarps of *Hippuris* have been reported in Western Europe only in the Pliocene and the Quaternary, whereas in Siberia they have been present since the Oligocene (Mai & Walther 1988). The Plioene–Quaternary endocarps of *Hippuris* have been assigned to *Hippuris vulgaris* L., whose endocarps are 1.3 × 0.7 to 1.9 × 1.0 mm in size. To our knowledge, three other species (morphospecies) have been described on the basis of Siberian fossil material: *Hippuris miocenica* Dorofoev (very similar to *Hippuris vulgaris*), *Hippuris parvicarpa* Nikitin and *H. minima* Dorofoev.

The small size of the Prüedo endocarps and the low length/width ratio indicates their affinity with the morphospecies *H. parvicarpa*, whose endocarps are, according to Dorofoev (1963), significantly larger (from 1.0 × 0.7 to 1.8 × 1.0 mm). The other morphospecies, *Hippuris minima*, has endocarps that are the same size (from 0.9 × 0.5 to 1.2 × 0.8 mm) as those of the Prüedo fossils, but are pear-shaped with a tapering base (Dorofoev 1963, figs 32 and 8–12).
In conclusion, despite differing from Hippuris minima Dorofeev and Hippuris parvicarpa Nikitin, the Prüedo fossils are identical in morphology and size proportion to the latter species. Although size may be a significant differential characteristic for the taxonomy of reproductive organs, the small number of specimens available does not provide a reliable statistical basis for assigning the Prüedo

Table 2. Pollen spectra of the sample Prüedo 4d, showing the total count of pollen grains and the pollen taxa

| Taxon Type            | Total Count |
|-----------------------|-------------|
| Taxodium type         | 1           |
| Myrica                | 11          |
| Sapotaceae            | 1           |
| Indeterminable Pinaceae | 8         |
| Mussaenda type        | 1           |
| Pinus                 | 44          |
| Rubiaceae             | 2           |
| Poaceae               | 106         |
| Microtropix fallax    | 1           |
| Plantago              | 1           |
| Distylium             | 1           |
| Amaranthaceae–Chenopodiaceae | 4     |
| Liquidambar           | 1           |
| Engelhardia           | 1           |
| Olea                  | 1           |
| Campanulacea          | 2           |
| Oleaceae              | 1           |
| Convulvulus           | 1           |
| Quercus 1–coccifera type | 2         |
| Rosaceae              | 1           |
| Quercus deciduous type | 7         |
| Caryophyllaceae       | 1           |
| Juglans               | 1           |
| Liliaceae             | 1           |
| Ulmus–Zelkova         | 1           |
| Apiaceae              | 3           |
| Salix                 | 1           |
| Cyperaceae            | 108         |
| Populus               | 1           |
| Indeterminable grains | 9           |

Table 3. Carpological specimens of the Prüedo deposits

| Family            | Taxon                  | Sample Prüedo |
|-------------------|------------------------|---------------|
| Cyperaceae        | Carex sp.              | 1a 4a 4b      |
| Alismataceae      | Alisma sp.             | *             |
| Ranunculaceae     | Ranunculus sp.         | *             |
| Labiatae          | Labiatae gen. et sp. indet | *     |
| Caprifoliaceae    | Weigela sp.            | *             |
| Hippuridaceae     | Hippuris cf. parvicarpa Nikitin | * |
| Linaceae          | Linum sp.              | *             |
| Rosaceae          | Potentilla sp.         | *             |
| Rubus sp. & Rubiella sp. | *                  |

*Taxon is present.

Fig. 8. Examples of the field appearance of the Prüedo facies association (FA) and the overlying glacial deposits. Location of the photographs within the stratigraphic columns is shown in Figures 6 and 7. (a) FA3, river flood plain to lacustrine environment (lignitic upper unit). The rhythmic alternation is made up of sandstone (1), siltstone (2) and lignite (3). (b) FA2, river flood plain and channel deposits (silty and micro-conglomeratic intermediate unit). The detritic layers in this part of the Prüedo sequence consist of conglomerate (1) and micro-conglomerate (2) alternating with siltstone (3). The clasts in the clast-supported conglomerate are sub-rounded and heterometric, from a few centimetres to tens of centimetres in diameter. The layers are tilted to the west and underlie a highly weathered decametre-sized block of granite in the till cover. A geological hammer, highlighted with a white circle, indicates scale. (c) FA1, river channel deposits (conglomeratic basal unit). Alternation of up to 2 m thick layers of conglomerate and siltstone. The clasts in the clast-supported conglomerate are rounded and are c. 10 cm in diameter.
material to a new species. Open nomenclature is therefore adopted. The discovery of *Hippuris cf. parvicarpa* Nikitin at Puëdro is especially interesting because it is isolated in the Pyrenees, and other records of similar plants have been reported in the Late Miocene only from northern regions (Siberia).

**Age constraints and palaeo-environmental conditions**

Our results are very similar to earlier palynological studies carried out in this basin (Jelgersma 1957) and in the Neogene infill of the Cerdanya Basin (Jelgersma 1957; Bessedik 1985; Barrón 1996a, b, 1997a, b, 1999; Pérez-Vila et al. 2001; Martín-Closas et al. 2005; Agustí et al. 2006; Sue & Fauquette 2012), confirming that the correlation made by Jelgersma (1957) between the Puëdro deposits and the Estavar deposits in the Cerdanya Basin is reasonable. The macrofossil remnants found in the assemblage in the Cerdanya Basin prompted Agustí & Roca (1987) to assign them to the Vallesian (continental stage between 11.1 and 8.7 Ma, according to Agustí et al. 2001). Thus, assuming that the Puëdro and the Cerdanya Basins are contemporaneous, the Vallesian is the best constraint for the age of the Puëdro deposits. A more conservative determination of the age of the Puëdro deposits is to assign them to the Late Miocene. The carpological assemblage is compatible with a Neogene age, and *Hippuris parvicarpa* Nikitin has been found only in deposits older than the Pliocene.

The pollen taxa identified in this study may be compared with the present-day plant ecosystems in southeastern China (Wang 1961), which is regarded as the closest living analogue of the Miocene South European flora (Sue 1984; Axelrod et al. 1996; Jiménez-Moreno 2005, 2006; Jiménez-Moreno et al. 2005, 2008; Jiménez-Moreno & Sue 2007). Therefore, in the Puëdro area the vegetation could be grouped into ecologically different environments depending on the degree of soil humidity and elevation. The environments range from low to higher elevations: (1) a broad-leaved rain forest and evergreen forest, from sea level to around 700 m, represented by *Taxodium* type, *Myrica, Distylium,* Sapotaceae, Oleaceae, *Mussaenda* type and *Engelhardia;* (2) an evergreen and deciduous mixed forest, above 700 m, characterized by deciduous *Quercus,* *Juglans* and *Engelhardia;* (3) above 1000 m, a middle altitude deciduous and coniferous forest with *Cathaya* and *Cedrus.*

The low representation of thermophilous elements and the higher representation of mesothermic taxa in the pollen spectra indicate that the climate in the Puëdro area was mainly temperate to warm-temperate. The Puëdro Basin was bounded by a helophytic belt consisting mainly of Poaceae and Cyperaceae and riparian elements such as *Salix, Liquidambar* and *Ulmus–Zelkova,* which are overrepresented in the pollen spectra at this site.

The carpological assemblage can be interpreted as a sporadic occurrence of a few aquatic (*Alisma, Hippuris*) or broad-ecology plants that grew very close to the deposition site. The occurrence of *Hippuris cf. parvicarpa* Nikitin in the Puëdro Neogene basin suggests that an isolated Neogene population of this water plant was present in the Pyrenees. Besides Puëdro, this plant has been found only in much higher latitude basins in Siberia. Such an isolated occurrence could be attributed to the Puëdro Basin being at significant higher palaeoaltitude in contrast to the lower altitude of all the other Neogene fruit- and seed-bearing basins in the Mediterranean region.

**Tectonosedimentary model for the Puëdro Basin**

The geomorphological, geophysical and palaeontological data show that the Puëdro sediments constitute the infill of a normal fault-related basin resulting from the negative inversion of the Gavarnie Thrust (the North Maladeta normal fault) at the end of the Miocene. As inferred from the audio-magnetotelluric survey carried out by Ortuño et al. (2008), the maximum depth of the basin at the toe of the fault is 200 m below the present-day surface, and decreases gradually towards the north along a 0.9 km transect (Fig. 5b). The additional data presented in this paper suggest that the palaeobasin had a lensoidal shape that was strongly controlled by the location of a fault stepover. As a result, the wider and deeper zone was located at the Puëdro transect, and the narrower and shallower zone at the Porèra transect. This survey and the surface geological mapping reveal the late superimposition of karstic processes at the eastern end of the basin (Figs 2, 3 and 5).
Besides controlling the geometry of the basin, the tectonic activity of the North Maladeta Fault might have played an important role in the distribution of the facies.

Figure 10a shows a sketch of the generation of the tectonic basin owing to the progression of faulting and tilting of the peneplain through which the Prüedo river system flowed. The regional peneplain formed before the onset of the North Maladeta Fault activity, as suggested by the presence of the peneplain at both sides of the fault. It might have been generated as an erosive surface before the emplacement of the Prüedo river system or as the result of the development of a wide alluvial plain, but in any case it reflects the absence of erosive rivers. The river feeding the Prüedo Basin was probably forced to migrate to the south as the movement of the North Maladeta Fault was tilted the peneplain, and thus, the alluvial plain developed on top of it. The preservation of the Prüedo sediments only next to the toe of the North Maladeta Fault suggests that the movement along the fault was contemporaneous with their deposition; that is, it suggests that (1) the fault generated the sediment trap, (2) the fault scarp acted as a tectonic barrier for the river flow and (3) the fluvial system ran parallel to the active trace of the North Maladeta Fault. The later deformation of the Prüedo sediments by the continuing activity of the North Maladeta Fault system cannot be excluded, but we believe that most of the tectonic displacement attributed to the North Maladeta Fault (up to 500 m cumulative vertical displacement) took place before the Quaternary. If the North Maladeta Fault system had not been active during the deposition of the Prüedo deposits but later on, we would expect a greater deformation on the Prüedo sediments, which are only locally tilted. Another feature that might indicate the synsedimentary activity of the fault is the general retrogradational pattern recorded in the stratigraphic sequence. This pattern reflects a decrease in the energy of the river system, which might be an autocyclic process, the indication of a change in the local base level and/or the result of the filling up of the basin related to a decrease in the North Maladeta Fault activity (i.e. a decrease in the generation of sediment accommodation space). At a smaller scale, the presence of repeated sedimentological cycles (sand–silt–lignite) in the uppermost part of the sequence could be an inherent fluvial feature and/or a climatic imprint. However, the tectonic control on these cycles should also be considered as an alternative cause. The tectonic control on the river flow capacity has been invoked by several researchers to account for sedimentological cycles in lacustrine and river systems in other parts of the Pyrenees (e.g. Martin-Closas et al. 2006) and elsewhere in the world (e.g. Ouchi 1985; Mukul 2000). In the Prüedo Basin, each tectonic pulse could have modified the geomorphology of the area, triggering changes in the river dynamics.

With respect to the palaeogeographical evolution of the area, several observations lead us to suspect that the river feeding the Prüedo Basin was captured by the Garona river (Figs 2 and 10). On one hand, the source area of the Prüedo Basin should have been coincident, or at least partially coincident, with that of the Garona river during the last glacial period. The Beret-Montgarri palaeovalley and the Bonaigua pass (Fig. 2), preserved to the NE–ENE of the Prüedo Basin and located at an elevation slightly higher than the Prüedo deposits (2000–2200 m), are the most probable sources of the Prüedo deposits as inferred from the paleoocurrent indicators and the lithological composition of the clasts observed. Although at present these areas are disconnected from the Garona river, at least the Beret-Montgarri palaeovalley was part of the drainage basin of the Garona river during the Last Glacial Maximum, as evidenced by the glacial striae on the polished surfaces found in this area and by the general slope of the planar valley base to the south (Zandvliet, 1960).

### Absolute uplift rates and recent exhumation histories: the role of the North Maladeta Fault

The thermophilous elements in the Prüedo deposits, like those in the Neogene deposits of the Cerdanya Basin, are rare. This contrasts with their abundance in the Catalan Coastal Basins (Fig. 1) during the Late Miocene (Bessedik 1985; Sanz de Siria Catalán 1993; Jiménez-Moreno 2005). In the Prüedo deposits, there is a better representation of vegetation characteristic of higher elevations such as *Quercus* (deciduous type) and conifers such as *Pinus*, *Cathaya* and *Cedrus* (Table 2). However, the absence of microthermic elements in the pollen spectra indicates that the area was situated at a lower palaeo-altitude than today. On the basis of the present distribution of plant species in SE China (Wang 1961), the vegetation identified in the Prüedo deposits would have been situated between 700 and 1000 m. Considering the present-day elevation of the sample Prüedo 4d used for the palynological analysis (c. 1890 m), this estimate implies an altitude change in the range of 900–1200 m since the Late Miocene. This altitude change is considered to be local surface uplift experienced by the North Maladeta Fault downthrown block, from where the pollen data originate. Because the throw of the North Maladeta Fault has been estimated at 440 m (offset of the peneplain at the Prüedo transect), the local surface uplift at the upthrown block has to be 440 m greater than that in the downthrown block (i.e. 1340–1640 m). This range of values allows us to estimate the regional uplift in the area since the Vallesian (11.1–8.7 Ma) as between 900 and 1640 m, which yields regional uplift rates between 0.08 and 0.19 mm a⁻¹.

These values of regional uplift need to be compared with data derived from independent sources. So far, no other data relating the post-orogenic surface uplift have been published for this part of the Pyrenees. The only data providing reference values for this region are the post-orogenic exhumation values obtained from fission-track studies. Fitzgerald et al. (1999) modelled the apatite fission-track data for three profiles in the Central Axial Pyrenees. Studying various sites along a cross-section of the Central Axial Pyrenees, those workers showed that the Maladeta Massif corresponds to the area where the maximum post-orogenic exhumation has occurred. This area experienced a 2–3 km exhumation during the last 10 Ma after the slowing down of the rapid post-orogenic exhumation. In other studies, Lynn (2005) and Sinclair et al. (2005) determined different post-orogenic exhumation histories for this area, but in both studies it was also concluded that the 2–3 km exhumation occurred in the last 10 Ma. More recently, Metcalf et al. (2009) combined the apatite data published by Fitzgerald et al. (1999) with additional biotite and K–feldspar fission-track data from the Maladeta Massif. The thermochronological data confirm a period of slow exhumation between 30 and 15 Ma ago, followed by a rapid exhumation of the massif in the Late Miocene–Pliocene (after 15 Ma ago). The comparison of the regional uplift data presented in this study and the exhumation histories established by the workers cited above (Table 4) shows that part of the 2–3 km recent exhumation reported could be explained in terms of regional uplift (0.9–1.64 km).

Besides the subregional data on exhumation, it is interesting to pay attention to the local variations in the exhumation values. The thermochronological studies of Lynn (2005), Sinclair et al. (2005) and Gibson et al. (2007) showed that the >2 km exhumation that occurred during the last 30 Ma in the Maladeta Massif (Maladeta profile) is greater than that obtained at sites less than 50 km to the north (Arties and Marimanhas profiles) (Fig. 2). The difference in exhumation, as deduced from the data by Gibson et al. (2007), ranges from 400 to 1300 m, depending on the site. This is in agreement with the Neogene activity of the North Maladeta Fault and
Table 4. Uplift and exhumation data for the Maladeta zone

|                      | Northern block of the North Maladeta Fault (km) | Southern block of the North Maladeta Fault (km) | Time range | Reference                          |
|----------------------|-------------------------------------------------|-------------------------------------------------|------------|------------------------------------|
| Uplift               | 0.9–1.2                                         | 1.34–1.64                                       | Since 11.6–5.33 Ma | This paper                        |
| Exhumation           | <2 km                                           | >2 km                                           | Since 29–30 Ma | Lynn (2005); Sinclair et al. (1999); Gibson et al. (2007) |
|                      | Subregional values                              |                                                 | Since 10 Ma | Fitzgerald et al. (1999); Lynn (2005); Sinclair et al. (2005) |

suggests that at least one part of the difference in exhumation is related to the North Maladeta Fault activity; that is, the difference in local uplift between the North Maladeta Fault hanging wall (Arties and Marimahans profiles) and the foot wall (Maladeta profile) (Table 4).

The regional uplift inferred in this paper for the Axial Zone of the Central Pyrenees has direct implications for the debate about the origin and preservation of the Pyrenean post-orogenic peneplains by Babault et al. (2005, 2007, 2009), Gibson et al. (2007) and Calvet & Gunnell (2008), among others. The 700–1000 m paleoaltitude of the Prüedo deposits can be considered as a minimum altitude for the peneplain remnants located at the base of the basin and surroundings, which suggests that this peneplain originated at an altitude lower than its present-day altitude (c. 2000 m). Such a paleoaltitude is much higher than the near-sealevel altitude proposed by Calvet & Gunnell (2008) for similar peneplain remnants in the Eastern Pyrenees, and much lower than the >2000 m altitude proposed by Babault et al. (2005, 2007, 2009) for the Pyrenean peneplain. In the light of the results obtained here, the more suitable explanation for the origin of the Pyrenean peneplains is that of Calvet & Gunnell (2008), who proposed that the erosional surfaces in a mountain belt might have a range of origins depending on both sub-regional and local history.

There has been a further debate on the feasibility of the preservation of such peneplains. On the basis of apatite (U–Th)/He ages, Gibson et al. (2007) concluded that the post-orogenic uplift of the Pyrenees has taken place at a constant rate for the last 30 Ma; however, this has been considered to be unfeasible because of the present-day preservation of pre-Quaternary high peneplains and the Quaternary rejuvenation of the landscape proposed by other workers (e.g. Babault et al. 2005). Gibson et al. (2007) also assumed a negligible uplift of this area during that time, as they interpreted the 2 km exhumation in terms of net erosion. The characterization of the peneplain surfaces in the Aran Valley performed by Ortuño et al. (2008) together with the dating provided here do not support the suggestions made by Gibson et al. (2007), Babault et al. (2009) and Metcalfe et al. (2009) and question the validity of the thermal model proposed by Gibson et al. (2007) by referring to incorrect assumptions about the location of the samples used.

The post-orogenic uplift rates obtained by various workers in the Axial Zone of the Western and Eastern Pyrenees are compared below with the 0.08–0.19 mm a−1 rates calculated for the Prüedo Basin and the Maladeta Massif. In the Eastern Pyrenees, the Cerdanya–Tét Fault system has affected a peneplain remnant, with vertical cumulative displacements that range from 500 to 600 m (Briais et al. 1990) since the Mid-Miocene. As a result, the Seu d’Urgell, the Cerdanya and Conflent Neogene Basins formed on its downthrown northern block (Roca, 1986, 1996). According to the anomalous palaeontological record of the Messinian terrigenous deposits, Pérez-Vila et al. (2001) estimated an uplift of c. 1000 m of the Cerdanya Basin since 6–5.3 Ma ago. Additionally, Calvet & Gunnell (2008) estimated ≥1000 m increase in altitude of the Cerdanya Basin during the last 12 Ma on the basis of pollen, fission-track and stratigraphic data. At some sites on the upthrown block, such as the Canigó Massif, the uplift could have attained 1500 m. Such values yield uplift rates of 0.08–0.12 mm a−1 during the last 12 Ma. More recently, Suc & Faquette (2012) have reported 0.06–0.12 mm a−1 uplift rates for the Cerdanya Basin on the basis of the pollen content of Miocene–Pliocene sediments. In the Western Pyrenees, in the carbonate Arbailles Massif, Vanara et al. (1997) and Vanara (2000) proposed 1000 m uplift since the Plio–Pleistocene and a 0.4 mm a−1 uplift rate since the Late Pleistocene on the basis of karst cavities, the fossil content of their infill, and U/Th dating. This greater uplift rate might be a local difference or the result of a greater uplift during more recent times as compared with the Central and Eastern areas.

To explain the post-orogenic uplift experienced in different parts of the Pyrenean range, isostatic compensation in response to Quaternary erosion and/or the loss of the subducted crust have been invoked (e.g. Calvet & Gunnell 2008; Gunnell et al. 2008; Ortuño 2008; Lacan & Ortuño 2012).

Conclusions

In the light of our findings, the following conclusions may be drawn.

1. The Prüedo Basin developed on top of a pre-existing peneplain and at the foot of the North Maladeta Normal Fault on the axis of the Central Pyrenees. The basin was lensoidal, c. 8.3 km long and less than 2 km wide. The infill consists of fluvial conglomerate, sandstone and siltstone coming from areas located between the north and the NE. The latest preserved records correspond to a palustrine system that favoured the formation of lignite layers.

2. The Prüedo deposits can be assigned to the Vallesian (11.1–8.7 Ma) on the basis of their pollen content and their correlation with the lignitic deposits of the Cerdanya Basin (Estavar, Sanavastre and Sansor localities).

3. The taxon Hippuris cf. parvicarpa Nikitin was identified for the first time in Europe, having previously been identified in Late Miocene deposits in Siberia. This and other carpalogical remains of Late Miocene taxa are consistent with the assigned age.

4. The age of the Prüedo deposits implies that the peneplain preserved in the area must have been formed before the Late Miocene.

5. The Prüedo palaeoflora indicates that the Prüedo Basin was formed at an altitude that ranged between 700 and 1000 m.
This estimate implies a local surface uplift of 900–1200 m in the North Maladeta Fault downthrown block and 1340–1640 m in the North Maladeta Fault upthrown block. When averaged since the Vallesian (11.1–8.7 Ma) these values imply uplift rates between 0.08 and 0.19 mm a^{−1}. These results are the first estimates of the post-orogenic uplift in the Central Pyrenees and are comparable with those of the Late Neogene uplift proposed for the Axial Zone in the Eastern Pyrenees.

These uplift values suggest that post-orogenic exhumation values obtained from thermochronological studies by other workers might partially reflect the uplift experienced in the area and not just net denudation. Exhumation values also reflect the activity of the North Maladeta Fault, as they are significantly higher in the upthrown block than in the downthrown one.

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