The Neogene Prüedo tectonic basin (Central Pyrenees)

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Abstract

The remnants of a palaeo-basin sedimentary infill outcropping at the Aran valley (Central Axial Pyrenees) have been mapped and studied combining stratigraphical and biostratigraphycal analysis with an audiomagnetotelluric surveying. The basin was developed on top of a preexisting peneplain and was formed by its offset as a half-graben related to the North Maladeta fault activity. The basin had an elongated E-W lensoidal shape and was infilled by an, at least, ~100 m thick sequence of fluvio-palustrine deposits derived from a palaeo-Garona river. The detritic sequence is composed by three major facies assemblages: a basal member dominated by conglomerates; a medium member mainly composed of sandstones and siltstones; and an upper member characterized by a rhythmic sequence of sandstones-siltstones and lignite layers. The results derived from the palynological and carpological analyses have allowed to constrain the age of the basin infill as Late Miocene, and very likely Vallesian; to characterize the vegetation of the belt surrounding the basin as a mainly temperate to warm-temperate assemblage and to estimate the palaeoaltitude of the site between 900 and 1000 m.a.s.l. Specimens of the taxon *Hippuris* cf. *parvicarpa* Nikitin have been identified for the first time in a European palaeoflora. The palaeontological dating of the basin infill allows estimating a minimum Upper Miocene age for the peneplains preserved in the area. The significance of the origin, preservation, and uplift of these landforms in the core of the Pyrenean orogen is discussed in the lights of the results of this work.

Keywords: palynology; carpology; audiomagnetotellurics; Neogene, North Maladeta fault; Central Axial Pyrenees

INTRODUCTION

The fluvio-lacustrine records are very valuable sedimentary archives of geological processes. Neogene lakes preserved in high mountain environments are especially relevant to the orogenic histories, since the generalized enhanced erosion during Quaternary times has left almost no witnesses of the geological history occurred between the end of the orogenic maximum and the emplacement of the Quaternary glaciers. The study of the Prüedo basin sedimentary infill at the core of the Pyrenean range has given the opportunity to partly decipher the environmental conditions prevailing during the Late Miocene in the area as well as to better constrain part of the tectonic history of the North Maladeta Fault (NMF), to which is closely related.

The Prüedo detritic sequence, in the proximity of the Aran Valley (Central Pyrenees, **Fig. 1, Fig. 2**), outcrops in two grassland plain areas at ca. 1900 m elevation that are "hanged" approximately 400 m above the current level of the river incision. In this high mountain environment, where almost no recent deposits have been preserved, the Prüedo sediments partly maintain their depositional geometry, which contrast with the highly deformed hard bedrock in which the basin was formed (**Fig. 2**). The post-orogenic (i.e., post-Oligocene) sedimentary cover has been almost completely eroded away by the emplacement of consecutive glacial and river systems during the Quaternary times. Besides the Prüedo sediments, and several yuxtaglaciar deposits generated during the Penultimate Glacial maximum (Bordonau, 1992), the only post-orogenic sediments described in the Central Pyrenees are those generated during or after the Last Glacial maximum.

In the present study, we describe two of the four outcrops of the Prüedo sequence observed in the area (**Fig. 3**). Site 3 and 4 have only been mapped, while site 1 (Riu Verd) and site 2 (Riu Merder) have been studied in detail by stratigraphic description and facies analysis, in order to determine the origin and provenance of the sediments. To constrain the age and the palaeo-environmental conditions of the deposits, some of the layers outcropping in site 1 and 2 have been analyzed with palaeontological purposes. The carpology and pollen content have provided with useful data, allowing updating the age and the interpretation of the origin of this sedimentary assemblage.

Pyrenean Neogene basins

The study area is located at the core of the Pyrenean range, part of a double north-south vergent Alpine orogen resulting from the collision of the Iberian and Eurasian plates (**Fig. 1**). The major convergent episode started in the Early Cretaceous and lasted until the Late Oligocene in the Eastern Pyrenees and until the Middle Miocene in the Western Pyrenees (Muñoz, 1992; Beaumont et al., 2000). After the end of the main collision, the Pyrenees have undergone a generalized period of stress relaxation with local episodes of compression. During this still lasting post-orogenic period, several intramountain basins were generated in the Axial Zone of the Pyrenees. To our knowledge, there is evidence of only three of these basins having been formed during pre-Quaternary times: the Arlas graben on the Pierre-St-Martín massif (Viers, 1977 as referred in Hervouët, 1997; Lacan, 2008) at the western part of the range; the Prüedo basin, aim of this study and located in the central part; and the Cerdanya-Conflent basin (Roca, 1986, 1996; Agustí and Roca, 1987; Cabrera et al., 1988; Clauzon, 2001; Martín-Closas et al., 2005; Calvet and Gunnell, 2008) at the eastern part of the Pyrenees. The sedimentary infill of the Arlas graben does not seem to have been preserved (Hervouët, 1997; Lacan, 2008). By contrast, in the Prüedo and the Cerdanya basins, ~100 m and ~500 m of sedimentary sequences have been preserved respectively, representing a unique record of the post-orogenic evolution and

Neogene palaeoenvironmental conditions of the chain. The palaeoaltitude and palaeoenvironmental conditions recorded in these basins infills are valuable data that might be combined with independent data to better understand the post-orogenic uplift histories of the Pyrenean range.

Geological and geomorphological setting

The remnants of the Prüedo deposits are located in the Central Axial Pyrenees, between the E-W trending Aran Valley, through which the Garona river flows, and the Aigüestortes-Maladeta massif (3404 m a.s.l. max. elevation). Basement rocks outcropping in this area consist of Paleozoic metasedimentary rocks intruded by Late-Variscan granitoids. As a result of the last Alpine orogeny, also Permo-triassic rocks are faulted and verticalized in the area. (**Fig. 1**). The Gavarnie thrust, one of the main Alpine structures on the Axial zone of the Pyrenees, was inverted in the post-orogenic period as a normal fault, and has been referred in the literature (Kleinsmiede, 1960; Bordonau and Vilaplana, 1986; Ortuño et al., 2008; Ortuño, 2008) as the North Maladeta fault (NMF). The relationship between the recent activity of this fault and the Prüedo sediments will be retaken up later on this work.

The Prüedo detritic sequence has been identified at four locations, three of them at the Prüedo plains and the westernmost one at the Porèra plains, all of them located less than 8 km north of the present E-W main water-divide of the Pyrenean range. These plains are located at approximately 2000 m a.s.l., at the watershed between the Aiguamòg and the Valarties valleys the former, and between the Valarties and the Ruda valleys the latter (**Fig. 2** and **3**). They are part of a group of high plains that have been interpreted as the remnants of a regional peneplain formed prior to the Quaternary glacial-river incision (Kleinsmiede, 1960; Zandvliet 1960; Ortuño, 2008; Ortuño et al. 2008). The S-N trending rivers incised in these plains are tributary headwaters rivers of the Garona river, which flows today into the Atlantic ocean. The Quaternary glacio-fluvial system has strongly modeled the area, so that, with the exception of the peneplain remnants, the present day geomorphologic features date from Quaternary times. Most of these features are erosive, and the only Quaternary sediments and landforms preserved today are post-glacial or glacial features from the Last Glacial period: a relatively thick (up to ~80 m) till preserved on top of some of the high peneplain remnants, some patches of lateral moraines and several rocky glaciers preserved in the valley sides (Bordonau, 1992; Serrat and Vilaplana, 1992). The post-glacial sediments are limited to debris-slope deposits in the valley sides, some alluvial sediments in the valley bottom and a several peat zones that have developed in the smoothest areas (**Fig. 2, Fig. 3**).

Previous studies of the Prüedo basin

The first detailed description of the Prüedo sediments was done by De Sitter (1954), although these had already been mentioned and interpreted as tertiary deposits by Volker (1953). In this pioneer report, De Sitter (1954) provided the stratigraphy of the Prüedo deposits and a basic mapping of three outcrops together with the basement lithologies, the glacial and recent sediments and the trace of the Gavarnie fault. In this mapping, the author delimitates the approximate extent of the Neogene basin in which the sediments would have been trapped. The proposed palaeobasin has an E-W elongated shape of ~5.5 km length and less than 1 km width. Later on, Jelgersma (1957) analyzed the pollen content of some of the Prüedo strata. The author ascribed a post-Vidobonian age (i.e., european continental Late Miocene stage) to the Prüedo sediments based on the comparison of the Prüedo pollen assemblage with that of the Estavar deposits (Cerdanya basin, Central-eastern Pyrenees). These data, discussed below in this paper, were taken into consideration in the regional works performed by Zandvliet (1960) and Keinsmiede (1960). Almost 30 years later, Vilaplana et al. (1986) retook up the study of the Prüedo sediments, describing in detail part of the stratigraphy observed in site 2 (**Fig. 3**). The authors also provided a sketch of the site location with respect to the Aiguamòg valley. Contrarily to the former works, performed by scientists of the Leiden School (Netherlands), which ascribed the deposits to the Tertiary, Solé Sabarís (1957) suggested that they could have formed in the last interglacial period. Additionally, Vilaplana et al. (1986) considered that alternatively to the Neogene origin, the deposits could also have been formed during the last glacial cycle and prior to the last glacial maximum as yuxtaglaciar lake deposits.

The sites mentioned by De Sitter (1954) are included in the cartographic synthesis of the Pyrenees made by Zwart (1965; 1979). Among the three outcrops considered in this cartography, the westernmost one has not been found in the present work. However, a different outcrop, located southwards of that one, has been identified and it is described in the following section.

Origin of the basin

Besides the different ages attributed to the Prüedo sediments, the works previously mentioned proposed contrasting hypothesis for the origin of the basin and its infill. As already mentioned, Vilaplana et al. (1986) suggested that the deposits could have been formed in a yuxtaglaciar lake in relation with the glaciar tongues flowing down the Aiguamòg valley. The scientists of the Leiden School, however, focused their interpretations in the relationship between the Prüedo deposits and the peneplain remnants, setting them in Tertiary times. De Sitter (1954) proposed that the sediments were the infill of a deep gorge incised in the original peneplain. On the contrary, Kleinsmiede (1960) argued that the sediments had formed prior to the peneplain because the top of the sequence coincided with it, and thus, the deposits were affected by the erosive process generating the peneplain. According to him, the sediments are possibly affected by the post-orogenic activity of E-W faults in the area, fact that would have favored their preservation. Also Jelgersma (1957) mentioned the tectonic imprint of the Prüedo sediments. However, she specified that the activity of the faults could have had a direct control on the origin of the sediments; the basin would have been the result of the surface subsidence by the activity of the faults. After its infill, the uplift of the area and the incision of the rivers would have been responsible for the present day configuration.

More recently, Ortuño (2008) and Ortuño et al. (2008) have provided new data supporting that the Prüedo sediments are the infill of a tectonic basin (the Prüedo basin) developed by the activity of the NMF. The activity of this normal fault after the main Alpine orogenic phase has been constrained by the combination of geomorphological, geological and geophysical methods, which have allowed the determination of the extension and depth of the Prüedo sedimentary basin in a S-N section across the Prüedo peneplain (**Fig. 3**). The results of this work indicate that the Prüedo basin was formed as a halfgraben at the NMF hanging wall (north wall) by the ~440 m offset of the old regional peneplain owed to the fault activity. The incision of the drainage network perpendicular to the NMF scarp has lead to the generation of tectonic faceted spurs that delineate the surface fault trace (**Fig. 3**). The comparison of the audiomagnetotelluric study and the geomorphological mapping performed by these authors suggests that the base of the basin corresponds to the peneplain palaeo-surface, which was tilted towards the south and compartmented in three tectonic grabens separated by secondary normal faults. In this research, the depth of the basin at the toe of the fault is constrained to be 200 m below the present day surface, and to decrease gradually towards the north along 0.9 km (**Fig. 4**). The stratigraphy of the observed deposits is interpreted as corresponding to a river system (river channel and flood plain) that is turned into a lacustrine system, as it is inferred from the fining-upwards trend and the presence of lignite layers at the top. In the following section, we describe in detail the stratigraphy observed in sites 1 and 2, updating the interpretation of the system recorded in the deposits.

MAPPING AND STRATIGRAPHY OF THE SEDIMENTARY INFILL

Four outcrops of the Prüedo detritic sequence have been identified in the area (**Fig. 3**). Only two of them, sites 1 and 2, are sufficiently large as to allow the general structure of the deposits to be distinguished, being sites 3 and 4 small patches of less than 10 m^3 grey layered siltstone. The interpretation of these last sites as belonging to the Prüedo *in situ* deposit is uncertain; since these sediments are located near the top of the peneplain surface and are surrounded by the till covering the peneplains, it is possible that they had been reworked and incorporated to the till as silt lenses embedded on it.

In addition to these outcrops, indirect evidence of the presence of the Prüedo deposits have been found in the Montaner creek (**Fig. 3**); here, the valley bottom morphology corresponds to long shaped lobular tongues of silt that could have formed by the downslope flow or creep of part of the Prüedo sediments outcropping in this site.

The stratigraphical description has been focused, thus, in sites 1 and 2. In each of these sites, several outcrops have been identified and named with an “N” or “S” letter and a number, depending on their location on the northern o southern slope of the creek. In sites 1 and 2, the outcrops are steep cliffs up to 40 m high. The area is affected by mass movements, which include mudflows and rotational landslides, which are inferred from the lobular morphologies, the former, and by semicircular scars, step like terraces and tilted beds, the latter.

The base of the deposits is not exposed in any of the outcrops. It is likely, though, that they are overlying the paleozoic basement. At site 2, the basal contact is not exposed but it has been inferred to be 3-4 m below the lowermost outcrop (S1), since the granitic basement outcrops at that position, down in the bed of the creek (**Fig. 3**).

On both sites, the uppermost part of the sequence is covered by a chaotic deposit, whose thickness ranges between 2 and 50 m. This deposit is made up of subangular granitic boulders with diametric sizes up to 3 m embedded in an arkosic matrix pale in color. This chaotic material is derived almost entirely from granitic areas at the S (Maladeta batholith), and has been interpreted as the till conformably covering the Prüedo sequence. In the particular case of outcrop S1 in site 1 (**Fig. 5 B**), the sequence is overlain by a ~7 m diameter block of weathered granite embedded in till. This site was not included in De Sitter (1953) previous mapping, which could be due to the fact that the site was covered by till and slope deposits before the 80’s, when a small dam and an artificial

channel were built, causing enhanced erosion to the head of the creek. During the course of this research, continuous landsliding on the slopes of this creek has provided new outcrops that have helped to better understand the nature of the detritic deposits. Thus, the previous work of Ortuño (2008) included the outcrop S2 in site 1 (**Fig. 3**) as part of the Prüedo sequence, which has been reinterpreted here as explained below.
The till covering outcrop S1 (site 1) seems to be part of a right-lateral moraine from the Rencules Glacier. Immediately north of this site, a small topographic plain located at the same height than the till top (1920 - 1940 m, 80-60 m above the valley bottom) is likely to correspond to the ridge top of this moraine (**Fig. 3, Fig. 5 G**). This till passes laterally to the E into conglomerates and siltstones with sandy lenses interbedded (site S1 in **Fig. 3**; **Fig. 5 H**) that, are likely to be yuxtaglaciar deposits, this is, to have been originated by glacial blockage of streams draining the Prüedo area during the Last Glacial maximum.

Stratigraphy

The sediments outcropping in site 1 are organized in layers that dip 20 - 30º to the E, while in site 2, layers are in subhorizontal (**Fig. 5 A**, **C**, **D** and **E**). In site 1, only 4.5 m of the base of the sequence are exposed. In contrast, site 2 has a total of 102 m thickness and represents a more complete record of the sequence. In this latter site, a fining upwards trend in the grain size is observed. This tendency was already noted by De Sitter (1954) and Jelgersma (1957). The beds can be grouped, according to the grain size of the sediments and the stratigraphic position, in three different units described as follows and summarized in the stratigraphic columns of **figure 6**.

Conglomeratic basal unit

The base of the sequence is mainly composed by metric (2 - 4 m) beds of heterometric and rounded grain-supported conglomerates with clasts < 10 cm diameter size. The clasts are very poligenic (quartzite-greywacke, schists, slates, and hornfels, with minor amounts of marble and granitic clasts) and have both planar and sub-spherical shapes. Some of the beds contain imbricated planar clasts, which indicate N-NE provenance. The conglomerate layers are interbedded with layers of clay-silt and layers of sand (medium and coarse grain size). The sandy layers present cross-lamination.

This assemblage is composed of Gm and Fl lithofacies (**Table 1**), as described in the classification proposed by Mial (1978), and seems to correspond to the deposit of a river channel, which locally migrates and is replaced by river lateral bars and flood-plain deposits.

The beds at this part of the sequence are compacted and locally cemented. This unit is best exposed in outcrop S1 in site 2 (**Fig. 3** and **Fig. 5 D**).

Silty and micro-conglomeratic intermediate unit

The intermediate unit outcrops largely in site 2, being the conglomerate layers better exposed in outcrop S1. It is composed of alternating layers of silt, sand, micro-conglomerates and minor conglomerates (**Fig. 5**. **A**). Some of the silty layers also contain lenses of sand, and locally, the unit is composed of 1-2 m thick layers of conglomerates. These latter are heterometric (up to 10 cm diameter) and have spherical and planar shapes, the latter displaying imbrications which indicate NNE and ENE provenance. The clasts lithologies are granite and schist, and more rarely, quartzite. They are grain-supported and locally, matrix-supported, with a sandy matrix. These conglomeratic levels have erosive bases, channel-like shapes and are not completely cemented. In the uppermost parts of this unit, the silt layers alternate with thin layers of lignite and are darker gray shade, which could be related to a higher content in organic matter.

In site 1, the assemblage outcropping N and S of the creek (N1 and S1) contains some organic rich silty layers interbedded with sandy lenses at the uppermost part and well cemented conglomerates at the lowermost part (**Fig. 5**. **B**). Since only 3-5 m of the sequence are exposed in these sites, it is not possible to determine if the layers correspond to the transition between the intermediate and the upper unit, or to a middle part of the intermediate unit characterized by interbedded organic rich material. These layers dip 20-30º to the E. As it has been mentioned before, they are covered by till, which suggests that the observed tilting could be due to glacial horizontal forces derived from the Rencules Glacier and/or till overweighting. Alternatively, the tilting of these layers could be linked to the neotectonic movement of the NMF or its secondary faults, as well as to a rotational landslide towards the Rencules valley. More evidence is needed to determine the cause of the deformation in this site.

This assemblage is composed of Gm, Fl and Ss lithofacies (**Table 1**) of the Mial (1978) classification, which could be interpreted as flood plain deposits and deposits of a river channel. Main differences with the 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, and 3) the lithologies of the clasts; a major representation of granitic lithologies and the absence of certain lithologies such as greywacke or hornfels.

Lignitic upper unit

The uppermost part of the sequence consists of alternating layers of grey silt and lignite, with occasional layers of medium and coarse sand. The layers are affected by 2 families of vertical orthogonal joints, likely to be the result of diagenetical processes (**Fig. 5 C**). The silt layers present two systems of orthogonal joints, with the vertical joints trending approximately N-S and E-W. The number of lignite layers observed in site 2 is 32. The lignite layers are up to 1 m thick and are gradually more abundant in the uppermost part of the sequence, where a rhythmic alternation, consisting of coarse sand, silt and lignite can be identified (**Fig. 5 E**). Five of these cycles have been identified in site 2. The sedimentological features of this unit fit the lithofacies description of Mial (1978) for types Fl and C (**Table 1**), which could be the record of a flood plain environment, temporally and repeatedly turned into a lacustrine system.

Changes in depositional environment

Site correlation and palaeoslope

The layers outcropping in site 1 cannot be correlated with those in site 2 because of their tilting towards the E. Since layers in site 1 might have been considerably displaced from its original position, which is unknown, they can not be considered in the reconstruction of the stratigraphy.

Layers in site 2, however, are approximately horizontal and undisturbed, and thus, they seem to be in their original depositional position. From the imbrications observed in site

2, it can be deduced that the detritic material was transported from the E-NE. No data about the palaeoslope has been found. At least for the medium and upper part of the sequence, where the layers are composed of finer grain sizes, the horizontality of the *in situ* layers can be assumed (**Fig. 3**), allowing to perform a correlation among the outcrops and to estimate the total thickness of the sequence, which reaches at least 102 m at this site (from 1839 m to 1941 m in altitude). This value is much smaller than the 200 m estimated by De Sitter (1954) and Jelgersma (1957), which was based in the altitude difference between the base of the outcrop documented (1820 m in site 2) and the altitude of the surface of the peneplain at this area (aprox. 2020 m). However, as it was deduced by the geomorphological and geophysical study (Ortuño et al. 2008), the peneplain in this area has not been preserved. The uppermost topographic plain is related to the surface morphology of the up to 30 m thick till deposits covering the Prüedo sediments. To estimate the maximum possible thickness of the original sediments, it should be considered that they were, in any case, as high as the peneplain on which the basin was formed. By extending the envelope of the preserved tilted peneplain surface to the south and by considering the base of the detritic sediments inferred from the audiomagnetotellurics survey (**Fig. 4**), this maximum thickness is ~300 m, about three times the thickness preserved up to today.

Facies interpretation

Previously, the Prüedo deposits had been interpreted as the record of lacustrine (Jelgersma, 1957) and possibly yuxtaglaciar fluvio-lacustrine (Vilaplana et al., 1986) systems.

As it has been commented in section 2, the Prüedo deposits are covered by till, and in site 1 (outcrop S1), also by yuxtaglacial fluvial-lacustrine deposits, exposed in outcrop S2. The lithologies and the roundness of the clasts in this fluvio-lacustrine assemblage suggest that the fluvial deposits come from pre-existing outcrops of the Prüedo deposits. Furthermore, the re-sedimentation of the Prüedo deposits is the most reliable explanation for the sequence exposed in outcrop S2 (**Fig. 3**; **Fig. 5 B**), since the drainage area (Prüedo plain and Salana's faceted spur) does not contain all the lithologies observed in the conglomerates and is too small as to provide the observed roundness of the clasts; Although the appearance of the silt and sandy layers is almost identical, the conglomerates in outcrop S2 are more matrix-supported and are embedded in a more silty matrix than the underlying conglomerates. A 10-15º tilt toward the W can be observed in some of the layers, suggesting that the assemblage could be the record of a small deltaic system formed in a yuxtaglaciar lake. The altitude range of the assemblage, aprox. 1808-1837 m, lower than the observed in the *in situ* outcrops of Prüedo deposits in site 2, supports the idea of this material being derived from the Prüedo deposits located to the E.

Outcrops N1 and S1 of site 1 and N1, S1-S5 of site 2 have been considered as part of the same system and named Prüedo deposits. The stratigraphic and sedimentological characteristics of the deposits are typical of an alluvial system grading into a palustrine system. The main features leading to this interpretation are the granulometric properties of the different assemblages, the facies association and the geometry of the beds observed. The alternation of grain supported conglomerates, sandstones and siltstones in the lower member is interpreted as the lateral migration of a fluvial channel in a fluvial plain. The facies association is also charateristic of fan-deltas, but the cross lamination observed in the sand layers indicates that this association did not correspond to such a system. The absence of coarse conglomerates in the intermediate member denotes a decrease in the energy of the system. This could be related to a switch in the position of the main channel or to a generalized decrease in the river capacity. If the fluvial plain was turned into a lake depression, the stagnant waters would be expected to have generated a drastic decrease in the river velocity, and thus, in the grain-size of the river load. A plausible explanation for the absence of coarse conglomerates, thus, is the transition from a river into a shallow lacustrine delta, which would be turned temporarily into a vegetated palustrine system represented by the lignites of the uppermost member. Further research, such as the analysis of the diatoms or ostracods content in the deposits of the intermediate member, would provide some insights relating the lacustrine character of the sediments. In any case, as it would be exposed in section 4, the pollen content of layers in the intermediate unit indicates palustrine conditions.

GEOPHYSICAL SURVEY

An audiomagnetotelluric (AMT) survey has been performed along a N-S profile in the Porèra peneplain (profile Munt, **Fig. 3**)**,** parallel to the section previously studied by Ortuño et al. (2008) at the Prüedo peneplain (profile Prüedo, **Fig. 3**; **Fig. 4**). The purposes of this study have been to check if there is evidence of the Prüedo deposits in depth and to determine the geometry of its expected confining basin. The profile has been performed on the downthrown block of the NMF.

The AMT method

In the magnetotelluric method, simultaneous measurements of the temporal fluctuations of the natural electromagnetic field are recorded at the earth surface, which can be related with the electrical resistivity of the subsurface (Simpson and Bahr, 2005). Variations in the resistivity values are related to changes in rock properties such as lithology, structure, porosity and fluid content. The audiomagnetotelluric method (Audiomagnetotellurics, or AMT) is a passive electromagnetic technique that works in the audio frequency domain and allows obtaining the resistivity distribution of the subsurface at depths from a few meters up to a few km.

Data acquisition, processing and subsurface modelling

This survey was performed using a *Stratagem* equipment from *Geometrics*, which works in the frequency range between 10^5 and 10 Hz. Data from a total of 11 sites were acquired along the North-South profile Munt, with a separation between contiguous sites of 50 m - 100 m.

The recorded time series were converted into the transfer functions (impedances) in the frequency domain using the Imagem Software (Geometrics, 2000) based on the Fast Fourier Transform and a least-square cross-spectral analysis. From the impedances, the apparent resistivity and phase were calculated at each site for different frequencies. The dimensionality of the data was tested using the Strike code (McNeice and Jones, 2001), showing an E-W directionality. Accordingly, the sites from the profile were projected along a NS line.

A 2D resistivity model along the projected line was obtained using the RLM2DI inversion code (Rodi and Mackie, 2001). The initial model consisted of a 100 Ω·m homogeneous halfspace, with a mesh of 98 (horizontal) x 58 (vertical) cells. Resistivity and phases corresponding to the TE and TM modes (transversal electric and transversal magnetic, respectively) were inverted, with an error floor of 10% for the apparent resistivity and 2.9º for the phases. The final model has a RMS value of 4.8. The comparison between the model responses and the data is shown in Fig. 6, along with the 2D model. Vertically, only the first half km is shown, given the loss in resolution at lower depths.

Geological interpretation derived from the resistivity model and field observations

By comparing the resistivity distribution model obtained and the surface geology, the following domains have been identified (**Fig. 7**):

"a"domain

The southernmost part of the section is formed by high resistivity materials (> 300 Ω·m) that extend from the surface down to, at least, 450 m depth, and correspond with the Tredòs granitic rocks outcropping in the Aiguamòg valley. Since the survey is located next to the intrusive contact between this batholith and Cambroordovician slates and quarzites, it is likely that the measured resistivity is an average value between the two basement lithologies. This domain is bounded to the N by a vertical discontinuity interpreted as a vertical fault known as the Tredòs fault, which limits the Tredòs granites with the Devonian limestones outcropping to the N.

"b" domain

This domain is characterized by low resistivity materials (10 - 80 Ω·m) located below the Porèra till cover (first 30 m of high resistivity material of 150 - 300 Ω·m). These deposits can be distinguished in two areas: the southern one, observed from 30 to 140-150 m depth, and the northern one, from close to the surface to 60 - 80 m depth.

The southern area is characterized by a planar base that dips gently towards the S. This surface could correspond to a part of the peneplain surface tilted by the tectonic movement along the NMF. Such a tilting is likely to be controlled by two tectonic structures: The eastern segment of the NMF, which bounds this domain to the S, and the Tredòs fault, that represents the northern limit of the southern area (**Fig. 7**). The proximity and dip of the NMF eastern segment with respect to the Tredòs fault suggest that both faults could be connected at depth. The eastern NMF displays a ~ 80º dip to the N, value that was also inferred for the central NMF (located 1 km to the S of this transect) in the Prüedo audiomagnetotellurics survey (**Fig. 4**).

At the northernmost zone, the base of the "b" domain has an undulating shape and might correspond with the remnants of the eroded peneplain surface. This area is where sites 3 and 4 are located (**Fig. 3**). At those sites, the Prüedo sediments seem to be part of lenses embedded in the coarser material of the till covering the area. The high content of granitic boulders forming this till could explain the slightly lower conductivity of this northernmost part with respect to the southern one.

In this way, only the southern area of this domain is identified as the Prüedo detritic sequence, while the northernmost area seems to be a distinctive basal unit within the till cover, and must contain significant amounts of reworked sediment from the Prüedo deposits.

"c" domain

The resistivity of this domain is moderate to low (10-15 Ω·m), and corresponds to the Devonian limestones outcropping in this area along a 50 m E-W band (**Fig. 2, Fig. 3**). The resistivity distribution obtained from this survey suggests that these rocks are present in a much larger extent at depth.

As in the Prüedo section (**Fig. 4**), the limestone unit in this part is sub-vertical and might reach more than 1.5 km in thickness. Sinkhole landforms observed at the surface (area between stations 04 and 05 in **Fig. 3** and **Fig. 7**) indicate that these rocks have been affected by karstification. The cavities are probably filled with a less resistant material than the till outcropping at the sinkhole top. To explain this difference in conductivity, two explanations are poposed here: either the karstic infill is exhogenic and is derived from the Prüedo detritic deposits or either it represents a concentration of conductive clay minerals in the sinkhole as a result of the formation of *terra rossa* derived from the limestones dissolution.

"d" domain

The "d" domain is characterized by resistivities greater than 300 Ω·m, corresponding to the Tredòs granitic rocks outcropping in the northernmost part of the Porèra section. Limestones from "c" domain are "embedded" by the Tredòs granite, which surrounds them both to the S and to the N. The irregular geometry of the northern contact of the Tredòs granite with the limestones suggests an intrusive rather than a tectonic nature for this limit.

Geometry of the Prüedo Neogene basin

The dimensions of the Prüedo Neogene basin are uncertain. A lensoidal shape basin in map view can be envisaged by looking at the geometry of the NMF trace and the location of the sites where the deposits have been preserved or inferred from the audiomagnetotelluric survey (inset in **Fig. 2**). The aerial maximum extension of the basin in a N-S direction, in the Prüedo and the Porèra transects, is indicated by the outcrop of the erosional planar landform at the northern part. This northern limit is located at 1,2 km and 1 km from the southern limit in the Prüedo and Porèra transects, respectively. However, the area where the most part of the sediments were trapped has significantly smaller dimensions: 0,5 km in Prüedo and 0,2 km in Porèra. At both sites, this depocenter zone is bounded by active tectonic faults.

The aerial extension of the basin seems to be the result of the step towards the south of the fault trace between the Aiguamòg and the Valarties valleys. Such a curvature in the fault plane might have caused secondary faulting and subsidence to accommodate the differential movement between the eastern and western fault segments, on the one side, and the central segment, on the other. The comparison between the information obtained from the two surveyed transects allows to do some estimates relating this matter; a tilting of the basin bottom is observed in both surveys. A greater tilt is observed in the Prüedo section (6 - 7º) than in the Porèra section (3 - 4º). In both cases, the tilting is greater than the observed in the larger distance profiles obtained by Ortuño et al. (2008) through the geomorphological analysis. This fact is in agreement with the development of a half graben geometry by the secondary faulting of the part of the hanging wall located next to the fault plane. An alternative genesis of a half graben by the roll over folding of the downthrown block has been discarded, since such a mechanism is not feasible in granitic rocks outcropping on the surface. Besides the greater tilting in the Prüedo section, the offset of the peneplain in it (440 m), as it can be derived from the geomorphological analysis performed by Ortuño et al. (2008), is smaller than the offset in the Porèra transect (515 m). This difference in offset also fits with a lower altitude of the basin bottom in the Porèra transect (~1750 m, **Fig. 7**) compared to the Prüedo transect (~1800 m **Fig. 4**). As it has been derived from the sedimentological features, the palaeoslope of the basin during the deposition of the lower member was towards the west. This fact contradicts the difference in altitude observed in the present and suggests that the lower altitude in the Porèra transect is the result of a greater tectonic subsidence and not a palaeotopographic feature of the fluvial basin. In the Porèra transect, the offset observed is likely the result of the sum of the tectonic displacement along 2 strands of the fault, which overlap in a narrow band (**Fig. 3**).

BIOSTRATIGRAPHIC ANALYSIS

In order to assess the age and the palaeoenvironmental conditions of the Prüedo sequence, several layers from site 2 (**Fig. 6**) were sampled and analyzed to determine the fossil content; one layer was sampled and processed for palynological purposes and three more layers were sampled and treated to separate and identify the carpological content. Besides these analyses, ~ 180 kg from three layers were sampled and sieved for a search of micro-mammals that only provided a piece of a phalanx and a dental piece corresponding to a canine tooth. Any of these pieces allowed identifying to which species they belong, so that the result proved to be insufficient as to motivate a larger sampling for micro-mammal fossils. A general inspection of the outcrops to search for fossils leaves of plants was also performed without success.

Palynological analysis

Method

The palynomorph-rich sample of this study comes from the lignite layer Prüedo 4d, located in site 2, outcrop 4S (Fig, 2; Fig. 5). In the chemical treatment, only 15 grams of sediment were used. The sample was processed with cold HCl (35%) and HF (70%), removing carbonates and silica respectively. Separation of the palynomorphs from the remaining residue was carried out using $ZnCl_2$ (density =2). Sieving was done using a 10 μm nylon sieve. The pollen residue, together with glycerin, was prepared on slides. A transmitted light microscope, using x250 and x1000 (oil immersion) magnifications, was used for identification and counting of palynomorphs. Even though palynomorphs are very abundant in the sediments, spores have not been considered due to their low representation.

The identification of the pollen grains was done 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, *Pinus* and indeterminable Pinaceae excluded, were counted in each analyzed sample (Cour, 1974).

Pollen assemblage

The sample is diverse, identifying a total of 33 taxa (Table 2). The pollen spectra analyzed indicate a flora dominated by Cyperaceae and Poaceae and other herbaceous taxa such as Amaranthaceae-Chenopodiaceae, Apiaceae, Campanulaceae, *Plantago,* Liliaceae, Caryophyllaceae, *Rumex*, *Convolvulus*, and Asteraceae. *Pinus* and indeterminable Pinaceae are relatively abundant and so is *Cathaya*, a mid-altitude conifer growing today in Southeastern 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 (**Table 2**). Some thermophilous elements occur and are represented by *Engelhardia*, Sapotaceae, *Mussaenda* type, *Distylium* and *Microtropix fallax*. *Quercus ilex-coccifera* type, *Olea*, Oleaceae and *Quercus* deciduous type have also been identified, the latter being relatively abundant. Besides *Cathaya*, the micro- and meso-microthermic elements, such as conifers indicating a high altitude (*Cedrus*, *Picea* or *Abie*s) are very rare and were only represented by *Cedrus*.

Carpology

Method

The three layers sampled for the carpological analysis correspond to fine homometric sand within well compacted layers, being all samples collected from site 2; one from outcrop S1 (Prüedo 1a) and two from outcrop S4 (Prüedo 4a and 4d). Approximately 1000 grams of sediment were processed following the procedure described in detail in Martinetto (2001, 2009); the sediment was broken up using a H_2O_2 3% solution. Part of the carpological remains floated to the surface and was collected using a cloth net of 1.5 mm spacing. The rest of the material larger than 3 mm diameter, was separated from the disaggregated sediment using a 0.3 mm metallic sieve, and was observed under the binocular lenses together with the material previously collected.

Results

The main result of this preliminary carpological analysis has been to point out that the Prüedo deposits can provide anatomically preserved carpological remains. These are, however, very scarce, certainly due to the small volume of sediment sampled and to the fact that sampling has been carried out by persons who did not have a particular search image for sediment types which could contain more abundant and better preserved fruits and seeds. Therefore, it would be advisable to plan further carpological researches in this basin.

The material recovered so-far consists in a few tens of mm-sized fruits and seeds from samples 1a, 4a, 4d, which are mostly deformed and partly abraded. This makes identification very difficult, also because very few carpological data have been published on the Neogene of the Pyrenees area. Nevertheless, the morphology of few tens of specimens was diagnostic enough to permit identification at genus level, by comparison to analogous material described in the literature or stored in the palaeocarpological collection of the Earth Sciences Department of Torino. The list of the identified carpological specimens is reported 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 and Walther, 1988).

One type of carpological remains deserves here a further treatment, being interpreted as endocarps of an extinct form of the aquatic herb *Hippuris*.

***Hippuris* cf. *parvicarpa* Nikitin**

Some of the endocarps of the specimens are nearly cylindric, ca 1.0 x 0.6 mm, 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 characterised by an oblique hilum plane (**Fig. 9**).

Fossil endocarps of *Hippuris* have so-far been reported in Western Europe only in the Pliocene and Quaternary, while in Siberia they are present since the Oligocene (Mai and Walther, 1988). The Pliocene-Quaternary records are referred to the modern species *Hippuris vulgaris* L., whose endocarps are 1.3-1.9 x 0.7-1.0 mm. To our knowledge, three other species (morphospecies) have been described on the basis of Siberian fossil material: *Hippuris miocenica* Dorofeev (very similar to *Hippuris vulgaris*), *Hippuris parvicarpa* Nikitin, and *H. minima* Dorofeev.

The small-size of the Prüedo endocarps and the low lenght:width ratio indicates their affinity to the morphospecies *H. parvicarpa*, whose endocarps are however significantly larger (1.0-1.8 x 0.7-1.0 mm) according to Dorofeev (1963). The other morphospecies, *Hippuris minima* has endocarps of the same size (0.9-1.2 x 0.5-0.8) than those of the Prüedo fossils, but they tend to be pear-shaped with a tapering base (Dorofeev, 1963, Fig. 32, 8-12).

In conclusion the Prüedo fossils differ from both *Hippuris minima* Dorofeev and *Hippuris parvicarpa* Nikitin, yet they are identical in morphology and proportions to the latter species. Size may be a meaningful differential character for the taxonomy of reproductive organs, however the small number of specimens available does not provide a reliable statistical basis to suggest assignment of the Prüedo material to a new species, therefore we adopt the open nomenclature.

The finding of *Hippuris* cf. *parvicarpa* at Pruedo is particularly interesting because it is much isolated in the Pyrenees area, and other records of similar plants are only reported from northern regions (Siberia) during Late Miocene.

Age constraints and palaeo-environmental conditions

We only have information coming from a single pollen spectrum. However, the results are very similar to previous palynological studies carried out in this basin (Jelgersma, 1957) and in the Neogene infill of La Cerdanya basin (Jelgersma, 1957; Álvarez-Ramis and Golpe-Posse, 1981; Bessedik, 1985; Barrón1996a, b, c; 1997a, b, 1999; Pérez Vila et al., 2002; Martín-Closas et al., 2005; Agustí et al., 2006) allowing to confirm that the former correlation made by Jelgersma (1957) among the Prüedo deposits and the Estavar deposits, in the Cerdanya basin is reasonable. The macrofossils remnants found in the assemblage from La Cerdanya basin lead Agustí and Roca (1987) to sit in the Vallesian (geologic period between 11.1 and 8.7 M.a., according to Agustí et al., 2001) the time of its deposition. Thus, if the Prüedo and the Cerdanya basins are contemporaneous, the Vallesian is the better constraint up to date for age of the Prüedo deposits. A more conservative determination of the age of the Prüedo deposits, however, is to assign them to the Upper Miocene. The carpological assemblage is compatible with a Neogene age, and *Hippuris parvicarpa* Nikitin has been only found prior to the Pliocene.

The pollen taxa identified in this study allow a reasonable comparison with the organization of the present-day plant ecosystems in southeastern China (Wang, 1961), which is regarded as the closest living analogue for the Miocene South European flora (Suc, 1984; Axelrod et al., 1996; Jiménez-Moreno, 2005, 2006; Jiménez-Moreno et al., 2005, 2007, 2008). Therefore, in the Prüedo area the vegetation could be grouped into ecologically different environments, which are arranged according to the degree of soil humidity and elevation (moving from lower to higher elevation environments) as follows:

1) A broad-leaved rain forest and evergreen forest, from sea level to around 700 m depicted 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 better representation of mesothermic taxa in the pollen spectra indicate that the climate in the Prüedo area was mainly temperate to warm-temperate. The Prüedo basin itself, independently of its palaeo-elevation, was bounded by a helophytic belt formed mainly by Poaceae and Cyperaceae and riparian elements such as *Salix*, *Liquidambar* and *Ulmus-Zelkova,* which are overrepresented in the pollen spectra from this site.

The carpological assemblage can be interpreted as a sporadic documentation of a few aquatic (*Alisma, Hippuris*) or broad-ecology plants which grew very close to the deposition site.

DISCUSSION OF THE RESULTS

Tectonosedimentary model

Source area

The conglomerates forming the Prüedo sequence could be derived from the wastage of other pre-existing detritic deposits in the surroundings, as it seems to be the case of the glacial fluvio-lacustrine deposits sitting on top of them. Two observations lead us to discard this secondary origin; on the one hand, there is no record of the hypotetical pre-existing conglomerates in the nearby area; on the other, the fact that the detritic sediments are well sorted is opposed to a re-sedimentation from a nearby site, which would generate more heterometric deposits. The conglomerates, thus, seem to be the direct record of the erosion and transport of bedrock outcropping in the area. According to the relationships proposed by Collinson, (1996), the general roundness of the clasts suggests that their original source area was, at least, several tens of kilometers away .Additionally, the lithologies of the clasts in the conglomeratic levels indicate that the material was transported from source areas located to the N. The southern provenance can be discarded since the areas located south of the Prüedo basin are composed, almost exclusively by granitic rocks, which are minor components of the conglomeratic clasts. This fact is in agreement with the N and NE flow directions that have been inferred from the imbricated clasts.

The drainage network at the end of the Miocene had to roughly match some of the palaeovalleys that are preserved today. The location of preserved peneplain patches at

2000 - 2200 m, that is, 400 - 500 m over the present day level of river incision, suggests two possible paths for the fluvial system headwaters; the palaeovalley located at Beret-Montgarri and the Bonaigua pass (**Fig. 2**), being the former one the most likely path, as explained below. These palaeo-valleys are the western and the southern limits of the Marimanhas massif, a granodioritic batholith sourrounded by paleozoic metasediments. These palaeovalleys are now connected with the drainage network feeding the Pallaresa river. However, according to the observations made by Zandvielt (1960), the Beret-Montgarri palaeovalley was connected to the Garona river during the Last Glacial maximum, as it can be inferred from the glacial striae in the polished surfaces found in this area and from the general slope of the valley planar-bottom towards the south. These features are not found in the Bonaigua pass, but this fact does not lead to dismiss that the area belonged to the palaeo-Garona network in Miocene times. Thus, the waters draining into the Prüedo basin could come from the Beret palaeo-valley, the Bonaigua pass or from both. The polygenic nature of the clasts composing the Prüedo sediments seems to better fit with the greater variability of lithologies outcropping in the Beret-Montgarri draining area (**Fig. 2**). Together with this observation, the palaeo-flow directions derived from the imbrictted clasts at different layers in the Prüedo deposits suggest the Beret-Montgarri area is the most probable main headwaters area of the palaeo-Garona river system. However, a detailed comparative study of the lithologies, including, for instance, their isotopic imprint, would be needed in order to better determine this matter.

Ciclicity of the upper unit

The upper unit of the Prüedo deposits is characterized by a rhythmic sequence of sandstones-siltstones and lignite layers. The thickness of the layers range from centimetric to decimetric and the lignitic member is lacking in some of the rythmites, especially at the lower part of the unit. This fining upwards cycles might have a climatic origin as well as a tectonic one. The climatic variations in the water discharge could be a simple explanation for the observed cycles; deposition of sand-size grains would be related with a greater energy of the flow, deposition of silty material with medium energy conditions, and the development of lignite would be favored by stagnant waters. Such changes in the alluvial system energy might reflect seasonal variations as well as temporarily larger climatic pulses. The tectonic control of lake level fluctuations has been evoked by different authors to explain this type of ciclycity in different lacustrine and river systems in the Pyrenees (e.g. Martín-Closas et al., 2006) and around the world (e.g. Laird, 1995; Ouchi, 1995; Mukul, 2000; Jain and Sinha, 2005). In the Prüedo sediments, the sedimentary cycles are likely to have a tectonic imprint since the active faulting of the NMF play a major role in genesis of the basin. With each tectonic pulse, the morphological changes in the basin might have caused changes in the river dynamics. The observed repetition (sand-silt-lignite), if having a tectonic origin, might be the result of sudden changes in the river flow due to the tilting of the alluvial plain or to sudden damping related to the generation of a tectonic barrier downstream and the subsequent installation of a palustrine system in previous channel areas. Such a tectonic barrier might correspond to one of the secondary faults located in the downthrown block of the NMF fault and to the W of site 2, this is, in the downwards sense of the flow (**Fig. 3**; **Fig. 10**). The different causes exposed above are difficult to test because the limited outcrops do not allow checking the lateral continuity of the cyclic layers.

Palaeolandscape

The Prüedo Neogene basin could be considered as part of a wider system, the Garona Neogene basin. It is very likely that prior to the formation of the Prüedo deposits (late Miocene times), the Garona Neogene basin was an endorheic system, disconnected from its present day outflow towards the Atlantic Ocean. Such type of isolated basins controlling the development of local erosional peneplains have been evoked by Calvet and Gunell (2008) as a possible explanation for the development of different disconnected peneplains along the Eastern Pyrenean range, such as those surrounding the Cerdanya basin.

The natural pounding of an endorheic Garona Neogene basin could thus be a simple explanation for the fining upwards character observed in the grain size of the Prüedo depositional sequence. In this case, this trend would be owed to a regional phenomenon, as it would be the case of the general maturing of the landscape. The setting of the present day ongoing river incision that caused the rejuvenation of the landscape in the Pyrenean range have been attributed to the capture of the Ebro palaeo-basin (Coney et al., 1996; Babault et al., 2005) in the Pyrenean southern foreland during the Pliocene, as well as to the global "hardening" of the climatic agents since the beginning of the Quaternary (Babault et al, 2005; 2006). It seems unlikely that the Garona river had ever been connected to the Ebro basin, and thus, the Quaternary river rejuvenation seems a more laudable explanation for the capture of the Garona basin by an atlantic river and the subsequent peneplain degradation.

An alternative cause for the decrease in the system energy could be the tectonic control of the NMF over the fluvial dynamics. Taking into account the formation of the peneplain excavated by the Neogene Garona river and the tectonic origin of the Prüedo basin, a tectonosedimentary model has been proposed here to explain the evolution of the sedimentary sequence. **Figure 10** shows a sketch of the generation of the tectonic basin by the progressive faulting and tilting of the peneplain through which the palaeo-Garona river flowed. The regional peneplain formed previously to the onset of the NMF activity, under the lack of erosive rivers and by the development of a probable wide alluvial plain dominated by a sinuous river system. In this smooth landscape, the tectonic tilting of the peneplain southwards by the NMF activity would have forced the river to migrate laterally and to run along the fault trace during periods of enhanced faulting. During these periods, the present day river path (following the axial trace of the Aran valley synform) was probably also active as a as depicted in **figure 10 C** (as a secondary channel or as the eroded neck of a tight meander). After the tectonic slowdown of the NMF, the river flow would have been switched to its Quaternary and present day path (**Fig. 1**). This switch of the river path should have occurred prior to the development of the Quaternary drainage network so that there has been enough time for the Garona river to entrench 400-500 m below the Prüedo higher preserved deposits into its present day path. This fact leads to think that most part of the tectonic tilting of the peneplain should have taken place between the Late Miocene and the end of the Pliocene at the latest.

Comparison with other Neogene basins in the axial Pyrenees

The Cerdanya Neogene basin (**Fig. 1**) and the Seu d'Urgell basin can be considered as contemporary analogues of the Prüedo basin. The two basins are intramontane tectonic basins located in the axial zone of the Pyrenean range, and have been filled by alluvial and fluvio-lacustrine systems during the Neogene. In both cases, the tectonic activity of the associated faults have offset the peneplain areas developed in these regions, leading to cumulative displacements that range from ~500 m for the NMF to 500-600 m at the Cerdanya-Conflent fault (Briais et al. 1990). The Cerdanya basin infill is also composed of conglomerates, sandstones and siltstones with lignite and thick lacustrine diatomite (Roca, 1986; 1996) that are likely to record a similar environment to the one recorded at Prüedo.

However, the most striking resemblance between the two sites is that both of them record a similar change in altitude since the time of the Late Miocene infill, issue that is discussed in the following section.

Several features make these basins differ one from each other. Whilst the Prüedo basin is related to a major normal component of faulting along the NMF, the Cerdanya has been interpreted as a extensional basin (Calvet and Gunell, 2008) but also as a transtensional basin related to right-lateral strike slip along the Cerdanya fault (Roca, 1986; 1996; Cabrera et al., 1988). Furthermore, the aerial dimensions of the Cerdanya basin are almost twice the Prüedo ones; for the Cerdanya basin, the geological record is considerably more extensive, since the sedimentary infill reaches more than 500 m. The existence of larger exposures have allowed to recognize three different tectonic stages (Roca, 1996); transtensional dextral slip followed by reverse faulting during the Miocene and the onset of extension during the Late-Miocene. With respect to the minor cyclic changes observed in the lignites of Prüedo, similar cycles were described in La Cerdanya by Martín-Closas et al. (2006) although the thickness of the layers is metric an not decimetric as the observed in the Prüedo upper unit, this difference could be explained in terms of a greater sediment load in the Cerdanya system.

Additionally, the wide sedimentary record of the Cerdanya and Urgell basins has a rich fossil content in flora and fauna, which has allowed establishing a detailed history of climatic, environmental and geographical changes along the Late Miocene (e.g.; Agustí et al. 2006; Martín-Closas et al., 2005). As already commented above in this paper, the resemblance of the pollen record of the two basins could be a good tool to correlate the Prüedo and the lignite deposits of the Cerdanya basin (Estavar, Sanavastre and Sansor localities). Although the data from Prüedo palaeo-flora is not complete enough as to confidetently compare it with the palaeo-flora of the Cerdanya, the pollen resemblance allows to make the assumption of a common age. However, an important fact lead to suspect that the Late Neogene Cerdanya basin was at a lower altitude than the Late Neogen Prüedo Neogene basin; the absence of *Hippuris* cf. *parvicarpa* Nikitin in the Cerdanya pollinic record seems to be indicative of a relatively warmer climate in comparison with that in the Prüedo area. Such a climatic difference still characterises these basins nowadays, since Cerdanya potential forest is dominated by *Quercus*, while the forest at the Prüedo basin is mainly made of *Pinus nigra*. Nevertheless, the present day difference in altitude between both basins, more than 800 m, is likely to be greater than the one during the Late Neogene, otherwise, their pollen contect would differ in a greater manner. In consequence, the uplift experienced at the Prüedo area should have been greater than the uplift at the Cerdanya.

The occurrence of *Hippuris* cf. *parvicarpa* Nikitin in the Prüedo Neogene basin is particularly interesting from the palaeobiogeographical point of view, because it suggests that in the Pyrenees area an isolated Neogene population of this water plant of northern

(Siberian) origin was present. At the light of other evidences, the reason of such isolated occurrence could indeed be attributed to a major palaeoaltitude of the Pruedo basin, in contrast to the low altitude of all the other Neogene fruit and seed-bearing basins.

Palaeo-altitude and differential exhumation of the area

The scarcity of thermophilous elements in the sample Prüedo 4d, only represented by the presence of Sapotaceae, *Microtropis fallax*, *Distylium*, *Mussaenda* type and *Engelhardia*, is common in the previous palynological studies from the Vallesian of La Cerdanya Basin (Jelgersma, 1957; Bessedik, 1985; Barrón, 1996a, b, c; 1997, 1999; Martín-Closas et al., 2005; Fauquette et al., 2006) and contrasts with the abundance of thermophilous elements in the vegetation during the Late Miocene vegetation of the Catalan coastal basins (Bessedik, 1985; Sanz de Siria Catalán, 1993; Jiménez-Moreno 2005). In Prüedo there is a better representation of vegetation typical from higher elevations such as *Quercus* (deciduous type) and conifers such as *Pinus*, *Cathaya* and *Cedrus* (Table 2). However, the absence of microthermic elements (i.e., altitude elements such as *Abies* or *Picea*) in the pollen spectra indicates that the area was situated at relatively low altitude, lower than its present day value of around 1800-2000 m a.s.l.. In contrast, relatively more microthermic elements such as *Abies* and *Fagus* were found in the Vallesian of La Cerdanya. The interpretation of an altitude lower than presently was proposed for La Cerdanya during the deposition of an overlying Messinian unit on the basis of comparison of the pollen spectra of this basin with the coastal Rosselló basin (Bessedik, 1985; Pérez Vila et al., 2001; Agustí et al., 2006). Following the present distribution of plant species in SE China (Wang, 1961; see above), the vegetation identified in the Prüedo area would be situated between 700 and 1000 m in altitude. However, the 700-1000 m palaeoaltitud estimated for the Prüedo deposits in base of the pollen results needs to be contrasted with data derived from different sources. Considering the present day elevation of the sample Prüedo 4d used for the palynological analysis (~1890 m), this estimate would imply a variation in altitude in the range of 900 - 1200 m since the Late Miocene. This increase in altitude does not take into account that the palaeo-altitude of the sea level during the Late Miocene could have been different from the present day one. Recent works concerning this issue (Müller et al. 2008; Kominz et al. 2008) provide values between 0 and 70 m with respect to the present day sea level for its global increase in altitude since 10 M.a. ago.

Local exhumation histories and absolute uplift rates

At this point, it is interesting to consider some of the data relating the post-orogenic exhumation history of the area inferred from fission track studies in the study area and surroundings; Fitzgerald et al. (1999) modeled the apatite fission track (ATP) data for three different profiles in the central axial Pyrenees. According to the results of this study, the Maladeta massif, located in the NMF footwall, corresponds with the area where the maximum post-orogenic exhumation has taken place. After the slowing down of the rapid post-orogenic exhumation between 30 and 20 M.a., this area would have experienced a 2 - 3 km exhumation during the last 10 M.a. More recently, Metcalf et al., 2009 have combined the apatite data published by Fitzgerald et al. (1999) with additional biotite and K-feldespar fission track data from the Maladeta massif. The thermochronologic data confirm a period of slow exhumation between 30 and 15 M.a ago, followed by rapid exhumation of the massif in the Late Miocene-Pliocene (after 15 M.a. ago). In other studies, Sinclair et al. (2005) and Lynn (2005) determined different post-orogenic exhumation histories for this area, but both works ciphered in 2 - 3 km the exhumation occurred in the last 10 M.a. Since part of the exhumation must be the reflect of the rock dismantling by erosion, only a portion of the 2 - 3 km exhumation of the Maladeta area must be the result of an absolute increase in altitude.

Based in 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 since the last 30 M.a., and consider unfeasible the up-to-today preservation of pre-Quaternary high peneplains and the Quaternary rejuvenation of the landscape proposed by other authors (e.g.: Babault et al. 2005). The characterization of the peneplain surfaces in the Aran valley performed by Ortuño et al. (2008) together with their dating provided here do not support the suggestions made by Gibson et al. (2007). Babault et al. (2009) also disagree with Gibson et al. (2007) and believe that their assumption of a recent outcrop of the sampled rocks might be wrong, and thus, it might invalidate the results of their model. Furthermore, Babault et al. (2009) consider that the increase in erosion during the Plio-Quaternary does not have much effect in the isotherms referred in the thermochronological studies, because it is concentrated in the narrow valleys and does not exceed, in any case, 1000 m. Metcalf et al. (2009) also objected the continuous exhumation since 30 M.a ago proposed for the area by Gibson et al. (2007), arguing that the elevation of the sample from were such a thermal model is derived might be too high as to record the late cooling episode.

In the particular case of the Maladeta area, the thermochronological studies performed by Sinclair et al. (2005) and by Gibson et al. (2007) showed that the values for the 30 M.a. exhumation occurred in the Maladeta massif (Maladeta profile) are greater than the exhumation obtained in sites sampled less than 50 km to the north (Arties and Marimanhas profiles) (**Fig. 2**). The difference in uplift range from 400 to 1300 m, depending on the site considered, as it can be deduced from the data provided by Gibson et al. (2007). This fact is in agreement with the NMF Neogene activity and suggests that the difference in exhumation between the NMF hanging wall (Arties and Marimanhas profiles) and the footwall (Maladeta profile) has a mainly tectonic origin.

With the available data, some estimates of the change in absolute altitude experienced by the Prüedo deposits can be done. By considering the offset observed today in the Prüedo transect (**Fig. 4**), a maximum of 440 m relative downward displacement can be expected to have occurred in the downthrown NMF block since the formation of the Prüedo deposits (Ortuño, 2008). This value needs to be considered as a maximum, since at the time of the deposition of the Prüedo sequence, the peneplain surfaces had to be already offset at least several decameters or even one hundred meters, becuase of the tilting and sinking of the peneplain was needed for the tectonic basin to be generated. But, which part of this offset is due to the downthrown movement of the hanging wall and which part to the upwards movement of the footwall?. The exhumation data commented above indicates that the two walls of the fault have undergone exhumation, this being greater in the upthrown block (southern wall). The range of 900 - 1200 m uplift since the Late Miocene inferred from the pollen data for the downthrown block (northern wall) has to be, at maximum, ~ 440 m smaller than the uplift of the upthrown block. Thus, a 1340 - 1640 m uplift for the Maladeta profile would have occurred since the Late Miocene (11,6 - 5.33 M.a.), indicating that more than one half of the 2-3 km exhumation obtained for the last 10 M.a. by different authors (see above discussion), would be the reflect of absolute uplift. The data inferred in this work allow, for the first time, to do a preliminary and rough estimate of the uplift rates of the study area since the Late Miocene, leading to values in the range of 0.08 and 0.22 mm/a for the NMF northern wall and 0.11 and 0.30 mm/a for the NMF south wall.

Palaeo-altitude changes in other parts of the chain

Presently, the data providing estimations of the changes in palaeoaltitude in the Eastern Pyrenees come from fission track (Calvet and Guinell, 2008), from palynological and other palaeontological studies (Pérez-Vila et al., 2001; Clauzon, 2001; Agustí et al., 2006; Calvet and Gunell, 2008) and from geomorphological and stratigraphical approaches (Clauzon, 2001; Babault et al. 2005; Calvet and Guinell, 2008). In this region, the Miocene palaeo-altitudes of the Conflent and Cerdanya basins (**Fig. 1**) have been discussed by different authors. Based on the location of the marine-continental transition and the present day topography, Clauzon (2001) estimates a minimum uplift of 200 m for the Conflent basin and a maximum uplift of 1300 m for the Canigó massif since the end of the Miocene. For the Cerdanya basin, different authors dealing with this subject obtain different results; Pérez-Vila et al. (2001) estimate that the uplift since 6 - 5.3 M.a. is ~ 1000 m, according to the pollen content of the Messinian infill. Additionally, the palaeontological record of the Messinian basin infill has lead to Agustí et al. (2006) to suggest that the palaeo-altitude was lower than the present day one. The authors conclude that the changes in elevation are partly controlled by the Cerdanya fault activity, occurring the grater uplift (at least 500 m) during the Vallesian, and a slower uplift of at least 1000 m after that time.

Another research carried out by Calvet and Gunell (2008) in this area estimate an increase in the palaeo-altitude of the Cerdanya basin ≥ 1000 m during the last 12 M.a.. In some higher zones, as the Canigó massif, the uplift could have reached 1500 m. This estimate is based on pollen, fission track data and stratigraphic data. According to the authors, the uplift is caused mainly by tectonism and, in minor account, by eustatic and climatic factors leading to enhanced erosion, and thus, greater isostatic uplift.

Less data are available for the exhumation histories of the Western Pyrenees. Vanara et al. (1997) studied a complex stepped system of karstic cavieties at the carbonatic Arbailles, which lead them to propose a 1000 m uplift of the area since the Plio-pleistocene. Massif. Vanara (2000) proposes a 0.4 mm/a uplift rate for this area since the Late Pleistocene, based on the morphosedimentary analysis, the analysis of the fossil content of the infill of these cavities, and the results of their U/th dating.

Causes of the recent uplift of the Pyrenees

Some final comments can be posed relating the causes of the Neogene-Quaternary uplift in the Pyrenees, issue that is under constant debate. The average erosion throughout the Pyrenees since the middle Miocene is around 500 - 670 m, according to the estimates made by Babault (2004). The author calculated the isostatic rebound of the lithosphere for such values of erosion following the model of Molnar and England (1990), obtaining a response in the range of 415 - 550 m of uplift, about half the values obtained here for the Prüedo area. Recent studies in the Alps (Schlunegger and Hinderer, 2001; Malusà and Vezzoli, 2006; Cadoppi et al., 2007) have shown that changes in erosion and tectonic activity from one valley to a consecutive one has an important effect in the distribution of the recent and present uplift rates. This fact leads to think that the average values obtained by Babault (2004) could be considerably greater in areas of enhanced erosion. The isostatic response to erosion is expected to be much smaller at the Pyrenees than at the European Alps, taking into account the smaller erosive capacity of the Pyrenean Quaternary glacio-fluvial systems. The values of tectonic uplift obtained for the Central Axial Pyrenees here (0.08 – 0.30 mm/a) are by far below the present day uplift rates referred by different authors for some areas of the European Alps, which range between 1,1- 1,6 mm/a and seem to partially reflect the isostatic response to Quaternary erosion (Champagnac et al. 2007; Schlunegger and Hinderer, 2001). At the southern Valais region, Champagnac et al. (2007) estimate in 0.5 mm/a the average isostatic rebound due to erosion since 1 M.a. ago. It is, thus, reasonable to consider than part of the uplift rates derived for the Prüedo area could be due to the isostatic response to Quaternary erosion. The isostatic response owed to the thermal reequilibrium (Muñoz, 2002) and the partial loss (Gunnell et al. 2008) of the subducted crust are other causes evoked to explain part of the uplift observed.

Among the main implications derived from this uplift are the genetical mechanisms and the preservation of the high altitude peneplains observed at different locations of the Axial Pyrenees, on one hand, and the palaeoflora and palaeofauna distribution, on the other. A generalized uplift of more than 1 km after the Late Miocene in the Central and Eastern Pyrenees is supported by the data obtained and discussed in this work. This fact questions the generation of the Pyrenean peneplain near its present day altitude proposed by Babault et al., (2005; 2006). Accordingly to these authors, the high altitude generation of the planar landforms during pre-Quaternary times is likely to be linked to the low energy of the rivers draining southwards into the Ebro basin. A dramatic dropdown of the base level as a result of the Ebro capture by the Mediterranean rivers would have caused river incision and peneplain degradation. Such an origin is not feasible in the case of the peneplain preserved in the Aran valley, unless the area had been connected to the Ebro draining network during the Neogene, which seems difficult to reconcile with the Prüedo palaeo-geography reconstructed here. The existence of other high altitude Neogene peneplain relicts in areas of the Western Pyrenees never linked to the Ebro basin (e.g., Lacan, 2008) suggests that different peneplain in different parts of the Pyrenees could have had different origins. Calvet and Gunell (2008) propose that the peneplains formed along the Pyrenees are more likely to be low-gradients piedemonts with different origins depending on the sub-regional and local history. Whichever is their origin, the results presented in this study support ~ 1 km uplift of the peneplain surfaces on top of which the Prüedo basin developed since the end of the Miocene. Future work needs to be done in order to clarify the amount of uplift that is involved in the exhumation of the chain and the possible causes of it. This includes, for instance, the search of palaeo-altitude indicators in other areas and the acquisition of present day uplift rates.

CONCLUSIONS

The results of the audiomagnetotelluric survey, the mapping and the stratigraphic and sedimentological analysis of the outcrops have allowed the geographical reconstruction of the Neogene Prüedo tectonic basin. The map view of the basin should have had a lensoidal shape, ~8.3 km long and less than 2 km wide, developed at the foot of the North

Maladeta fault as the result of the extension associated to it. This tectonic low was filled by fluvial conglomerates, sandstones and siltstones coming from areas located to the N-NE. The system latest preserved records correspond to a palustrine system that favored the formation of lignite layers. The infill of the basin shows a gradual fining upwards tendency that might be related to 1) the natural pounding of the system in relation with the maturing of the landscape or to 2) the lateral migration of the main channel towards its present day path as a result of a slowing down of the NMF activity.
The review of the correlation between the Prüedo and the lignitic deposits of La Cerdanya basin (Estavar, Sanavastre and Sansor localities) on the basis of their pollen content points towards the Vallesian (11.1 - 8.7 M.a.) as the most probable age for the Prüedo infill. This result is reinforced by the identification of carpological remains of Late Miocene taxa. The taxon *Hippuris* cf. *parvicarpa* Nikitin has been identified for the first time in Europe, being the previous identifications located in Late Miocene deposits of Siberia.

Together with the age of the basin, the palaeoenvironment and palaeoaltitude of the basin inferred from the palaeontological analysis provide new insights for the pre-Quaternary history of the Central Axial Pyrenees. The Prüedo palaeoflora witness the existence of a warmer than today climate during the Late Miocene and suggests that the area could have experienced up to 1 km uplift since that time. This data is in agreement with the exhumation histories preposed by different authors for this area as well as with the values for the Late Neogene uplift proposed for other parts of the chain. The results of this study imply that the planar landforms on which the Prüedo Neogene basin is sitting originated prior to the Late Miocene and, most probably, at an altitude below 1000 m.a.s.l..

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Figure Captions

Figure 1. Location of the study area in the Pyrenean range. The Neogene intramountain basins in the Axial zone are located: 1, Arlas; 2, Prüedo; 3, Urgell; 4, Cerdanya-Seu d'Urgell; 6, Conflent. Also the Canigò massif is located (5). The main geological units are shown: a, Paleozoic basement; b; Mesozoic cover; c; Synorogenic Tertiary cover; d; Neogene and Quaternary cover. The main onshore active faults are maked in bold.

Figure 2. A.Geological map of North Maladeta Fault (NMF) and its sourroundings. Legend: 1, Gavarnie thrust; 2, North Maladeta Fault; 3, triangular facets; 4, upthrown peneplain; 5, downtrhown peneplain 6, glacial till and slopedeposits; 7, Neogene Prüedo deposits; 8, Permo-triassic metamorphic lutites and sandstones; 9 Carboniferous limestones and detritic rocks with variable degrees of metamorfim; 10, Devonian limestones and detritic rocks with variable degrees of metamorfim; 11, Silurian slates; 12; Cambro-ordovician sandstones, schists and slates; 13, Maladeta granodioritic rocks; 14; Other granitoids; 15; volcanic rocks. The areas in white correspond to the Late Quaternary sediments such as fluvial sediments and alluvial cones. Some of the lithologies can only be distinguished in Fig. 3. **B**. Sketch with the hypothetical dimensions of the Neogene Prüedo basin. The peneplain remnants preserved today are indicated in dark gray (uplifted) and light gray (downthrown).

Figure 3. Detailed map of the study area. Location of the four sites where the Prüedo sequence outcrops are shown. The stations surveyed with the audiomagnetotellurics method in the Porèra and the Prüedo transect have been also marked. The most relevant geomorphological features for the study of the Neogene basin have been included. See Fig. 2 for the legend of the lithologic units.Legend: 1, slided material; 2, alluvial cone; 3, rocky glacier; 4, sinkhole; 5, fluvial terrace; 6, peat areas; 7, outcrops of Neogene Prüedo deposits; 8; North maladeta fault; 9, normal fault scarp;10; landslide crown scar.

Figure 4. Geologic interpretation of the Prüedo section magnetotelluric survey (located in Fig. 3), modified from Ortuño et al. 2008. The Prüedo basin is envisaged in this section as formed on top of a preexisting planar surface by the activity of the North Maladeta Fault and secondary faults. The maximum extension of the Neogene basin as derived from this section is 1.2 km.

Figure 5. Examples of the field appeareance of the Prüedo deposits and the overlying yuxtaglaciar deposits. Whenever possible, the location of the pictures within the stratigraphic columns has been shown in Fig. 6. **Outcrop S1, site 2**: **A)** conglomeratic layer alternated with siltstones; **D)** conglomeratic layer. A pen was used as a scale; **Outcrop S4, site 2**: **C)** orthogonal jointing in the lignite layers; **e)** rhythmic alternation made up of sandstone (1), siltstone (2) and lignite (3); **Outcrop S1, site 1**; **B)** tilted detritic layers from the Prüedo sequence underlying a highly weathered decametric block of granite in the till cover. A geologic hammer was used as a scale; **Outcrop S2, site 1; H)** lateral transition between the Quaternary till and the yuxtaglacial detritic sequence overlying the Prüedo deposits;

Figure 5, continuation. F) conglomerates within the yuxtaglacial detritic sequence shown in f). The Prüedo peneplain viewed from the peneplain to the W (**G**). The lateral moraine top corresponding to the Rencules hanged glacial valley, the faceted spur of the NMF scarp below Salana's peak and the site 1 has been indicated for their location.

Figure 6. Stratigraphic columns of sites 1 and 2. Include sample's name and sample's destination. Some of the pictures in Fig 5 are located for reference.

Figure 7. Porèra Section. **A**) Resistivity model obtained from the 2D inversion of AMT data. **B**) Pseudosections of the measured phases, TE and TM modes. **C**) Pseudosections of the model phase responses, TE and TM modes.

Figure 8. Porèra section. Resistivity model (above) and geological interpretation (below). Resistivity gradients observed in the model are marked with lines. White lines correspond to lithological changes interpreted as non-tectonic (erosive or intrusive) contacts. Red lines are tectonic contacts corresponding to the main fault (NMF), the Tredós fault and secondary conjugate faults. The geological section is derived from the integration of AMT data and field survey.

Figure 9. Photos of the seeds of *Hippuris* cf. *parvicarpa* Nikitin found in samples *Prüedo* 1a and 4b taken with a binocular lens. Norte the low degree of preservation of the seeds in these samples. The scale on the left corresponds to one milimiter grid.

Figure 10. A) Picture of the Prüedo peneplain below the Salanas peak triangular facet (located in **fig. 2**); **B)** Block diagrams showing the paleogeographic evolution of the Prüedo area during the Neogene: **1)** The present-day highest peaks of the area were *inselbergs* standing out within the peneplain through which the Neogene Garona river flowed prior the Prüedo basin formation; **2)** The tectonic activity of the NMF leads to the

formation of a semi-graben in the downtrhown fault block and to the tilting southwards of the peneplain, which forces the river to migrate towards the fault foot; **3)** Filling of the Prüedo Neogene basin takes place during the Upper Miocene. The maximum thickness of the deposits is confined to tectonic subgrabens north of the Salana's peak area; **4)** Presently, the landscape of the area allows to recognize remnants of the offset palaeo-peneplain, patches of the Prüedo detritic infill and the triangular facets that bear witness of the NMF Neogene activity.