1. Introduction

1.1 Rationale

Research over the last two decades into the evolution of mountain systems has focussed on the controls and mechanisms of orogenic growth and the interactions of tectonic and surface processes in shaping the mountain landscape (e.g. Beaumont et al., 1992; Koons, 1989; 1995; Willett et al., 1993). These studies have highlighted the interactions of controls such as tectonics and climate and their relative effects on the overall evolution of mountain belts (Willett, 1999; Beaumont et al., 2000a). However, these studies have mainly concentrated on active orogenic systems and have not considered the effects on post-orogenic systems. Recent research has started to redress this imbalance (Pazzaglia and Brandon, 1996; Baldwin et al., 2003), yet the relative importance of the various controls on the resultant geomorphology or the record of erosion in post-orogenic systems has not been well documented.

This study focuses on the post-orogenic evolution of the Pyrenean mountain belt, by applying digital elevation data, field observations and low temperature thermochronology. These techniques are used in order to examine the controls on the macrogeomorphic form and to quantify the degree and timing of post-orogenic exhumation and erosion.

Previous work on the Pyrenees by Morris (1999) documented the gross asymmetric form of the orogen with a steeper retro-side and corresponding gentler pro-side. The pro-side of an orogen may be defined as that above the subducting plate, whilst the retro-side is defined as that above the overlying plate (figure 1.1). This thesis builds upon these observations, by using digital elevation data to characterise the macrogeomorphic form of the Pyrenees at a catchment scale. This provides a mechanism by which to improve the understanding of the controls on the post-orogenic system.

Prior low temperature thermochronological research by Gibson (2005) in the central Pyrenees is complemented by new data acquired during this study. Observation and modelling of this combined dataset allows quantification and timing of exhumation and erosion and also allows the transition from a syn- to post-orogenic system to be documented and analysed.
1.2 Background to research

1.2.1 Introduction and definitions

Orogenic belts form at the margins of tectonic plates as a consequence of collision (Dewey and Horsfield, 1970). In this study an active orogen is defined as one in which convergence is still taking place and the plates are still moving towards one another. In contrast, a post-orogenic system is defined as one in which the active convergence has ceased and the two plates are no longer moving with respect to one another.

1.2.2 Active orogenic systems

The development of active orogens can be approximated using the concept of critically tapered wedges (Chapple, 1978; Dahlen, 1990; Dahlen and Suppe, 1988; Davis et al., 1983; Platt, 1986; 1993; Willett, 1992). This theory is based on the consideration of mountain belts as slabs of Coulomb type material that strive to maintain a critical angle of taper at all times (figure 1.1). This was proposed as a result of the observation that wedges in modelling experiments approximate the forms seen in mountain belts, whilst maintaining a constant angle of taper at all times (Davis et al., 1983). Wedges grow by one of two processes: either by frontal accretion, with material incorporated at the toe of the wedge or by underplating or basal accretion, where material is incorporated into the wedge at depth. The balance between the processes of frontal and basal accretion strives to maintain the critical angle of taper at all times (Davis et al., 1983). If the angle becomes too great then the wedge will begin to lower the angle, by concentrating the focus of accretion to the toe of the wedge. Conversely if the angle becomes too shallow, then basal accretion will dominate in order to increase the critical angle, to restore balance to the wedge. Other features such as out-of-sequence thrusting, backthrusting and ductile deformation also play roles of varying importance in acting to re-equilibrate the wedge (Platt, 1988). The critical angle maintained is dependent on a number of factors including erosion rates, isostasy, changes in basal friction and wedge rheology (Davis et al., 1983; Platt, 1986; Koons, 1990; Willett, 1992; Willett et al., 1993). Estimates for the critical angle from both modelling experiments and measured natural examples are on average within the range of two to five degrees (Davis et al., 1983).
Mountain belts are not however, one-sided, and as such they can be represented by two critically tapered wedges back-to-back to one another. The nomenclature by which these wedges are differentiated is dependent on the polarity of subduction of the mountain belt. The wedge and associated foreland basin above the subducting plate are denoted by the pro-prefix, and those above the overlying plate by the retro- prefix (figure 1.1). The position where the subducting plate meets the overlying plate is known as the interplate convergence point, singularity or S-point (Willett et al., 1993). Growth of the two wedges during orogenesis leads to the distribution of material in such a way as to maintain the critical angle of taper in both pro- and retro-wedges. A generalised evolution for the growth of mountain systems is given in figure 1.2 (Willett et al., 1993). Modelling work has shown that following the development of initial uplift, both the pro- and retro-wedges begin to grow. The lines on figure 1.2 represent instantaneous flow lines, showing the general direction of motion of material during the growth of the mountain belt. The majority of material is accreted into the pro-wedge, with the retro-wedge effectively pinned, with little accretion occurring, related to the majority of material being removed from the subducting plate. This leads to a general flux of material from the pro- to the retro-side and the development of an asymmetric form. The pro-wedge in this system exhibits a minimum critical angle of taper in contrast to the retro-wedge which exhibits a maximum critical angle of taper (Dahlen, 1984; Willett et al., 1993). The respective geometries of the pro- and retro-wedges limit the development of a foreland fold and thrust belt on the retro-side, whilst promoting the development of a belt on the pro-side. The combination of frontal and basal accretion in these wedges leads to the critical angles being maintained.

The development of an asymmetric form for a mountain belt is predicted for all scenarios in which there is a component of frontal accretion (Willett et al., 2001). This observation from modelling studies is consistent with that observed in many active orogens around the world, such as Taiwan and the Southern Alps of New Zealand. The position of the main drainage divide, defined as the upper boundary between the pro- and retro-wedges, is found approximately above the S-point of the orogen (Willett et al., 1993). If tectonic forces alone are considered, then the position of the drainage divide is predicted to fluctuate only slightly above this point as thrust sheets are incorporated into the wedge (Naylor, 2004).

These observations on the development of wedges described so far have concentrated solely on the tectonic controls on the form of the wedge. As mountain belts grow in elevation, they are likely to cause perturbations in the local climate. In particular they have the potential of
developing enhanced orographic precipitation on their windward flanks and a corresponding rain shadow on their leeward side. It has been suggested that enhanced precipitation and hence greater discharge on one side of an orogen, may result in a change in patterns of erosion and cause the principal drainage divide to be shifted within the orogen (Willett, 1999; Roe et al., 2002; 2003). This migration is determined by the relative efficiency of both the horizontal tectonic forces and the degree of erosion acting against them.

The coupling of tectonic and surface process models has sought to consider the relative degree of importance of the tectonic and climatic forces in the development of the form of mountain systems. One of the principal conclusions from these studies is that tectonic forces dominate in active systems, with the action of horizontal tectonic motion being the principal control on the macrogeomorphic form of mountain ranges (Willett et al., 2001; Burbank et al., 2003).

1.2.3 Post-orogenic systems

In contrast to the vast amount of research carried out on active orogenic belts, considerably less work has examined the possible controls on the evolution and development of inactive orogens. These orogens provide a perplexing macrogeomorphic problem of the persistence of topography remaining from orogenic events that ceased millions of years ago (Pazzaglia and Brandon, 1996). In active systems, the principal control on the macrogeomorphic form appears to be tectonics, with the horizontal compressive forces driving the orogen to develop an asymmetric form (Willett et al., 2001; Naylor, 2004). However, the relative importance of controls in the post-orogenic system are not as well understood. Three controls are considered to be of importance in the post-orogenic system: tectonics, climate and lithology (Pazzaglia and Brandon, 1996). Their interactions and the degrees of relative importance have been debated in the post-orogenic evolution of ranges such as the Appalachians (Hack, 1980; 1982; Barron, 1989), with no clear conclusions reached on the roles they have played in shaping the landscape (Pazzaglia and Brandon, 1996).

Models of river incision in post-orogenic systems has shown that there are a number of factors that encourage the persistence of topography over time. These include lithology, climate, syn-orogenic uplift rates, channel bed alluviation, flexural isostatic response and the magnitude of a critical shear stress for channel bed erosion (Baldwin et al., 2003). A

Chapter 1 - Introduction
combination of these factors was shown to allow topography to persist for hundreds of millions of years, explaining its persistence in old inactive orogens.

The Pyrenees is such a post-orogenic mountain belt, with tectonics having ceased in the mid Miocene. Compared to examples such as the Appalachians it is relatively youthful recording an approximately 20 Myr post-orogenic history as opposed to a 250 Myr post-orogenic history. Therefore, in addition to looking at the post-orogenic system, we also have the opportunity to analyse the transition from a syn- to a post-orogenic mountain belt.

### 1.2.4 Pyrenean mountain belt

The Pyrenean orogen may be defined as an asymmetric doubly-vergent mountain belt formed by the collision and partial subduction of the Iberian Plate beneath the European Plate (ECORS Pyrenean Team, 1988; Choukroune et al., 1989; Roure et al., 1989; Muñoz, 1992; Verges et al., 1995; Beaumont et al., 2000b). The relatively small size of the orogen, quality of exposure, preservation of the synorogenic deposits and the volume of geophysical data that has been collected on it provide a good example from which to study the processes of orogenesis (Beaumont et al., 2000b; Vergés et al., 2002). The Pyrenees form an east-north-east, west-north-west topographic barrier between France and Spain, extending from the Bay of Biscay in the west to the Mediterranean Sea in the east (figure 1.3). Relatively uniquely, the Pyrenees show no documented evidence for metamorphic or plutonic processes associated with their Alpine collisional history. As a consequence, evidence from metamorphic and igneous events dating from the earlier Hercynian Orogeny are well preserved (Muñoz, 1992).

The boundary between the Iberian and European plates has had a long and varied history, recording evidence for both compression and extension at different stages in its evolution (Zwart, 1979; Vergés et al., 2002). This section reviews the history of the Pyrenean mountain belt and the principal geology relating to each stage in its development.

The modern Pyrenean system can be divided into a series of tectonic or structural units (figure 1.4). These units form the basis for descriptions of localities used throughout this thesis. Six units may be defined, from north to south (Muñoz, 1992; Vergés et al, 1995):
1) Aquitaine Retro-Foreland Basin

2) North Pyrenean Thrust Belt

3) North Pyrenean Fault Zone

4) Axial Zone

5) South Pyrenean Thrust Belt

6) Ebro Pro-Foreland Basin.

The location of these units is shown in figure 1.4. The following section examines the evolution of the Pyrenean mountain belt and includes descriptions of the above units in their geological context.

### 1.2.4.1 Pre-collisional history

#### 1.2.4.1.1 Basement and Hercynian tectonics

The basement of the Pyrenean mountain belt comprises Palaeozoic aged rocks that outcrop within the Axial Zone. The lower part of this basement can be divided into five main lithological groups (Zwart, 1986):

1) A sequence of high-grade metamorphic rocks, which may represent a pre-Hercynian basement

2) Palaeozoic sediments, with accompanying minor volcanics

3) Metamorphosed Palaeozoic sediments

4) Intrusive Granodiorites

5) Pre-Hercynian Orthogneiss.
Some of the North Pyrenean massifs (Agly, St. Barthelemy, Castillon) are suggested to be a basement to later Palaeozoic rocks, with disputed ages ranging from Pre-Cambrian to Early-Cambrian (Zwart, 1979; 1986). They are composed in part of Granulite facies rocks derived from sediments and include some charnockitic intrusives (Zwart, 1979; 1986; Vissers, 1992).

The deepest exposed sedimentary rocks in the Axial Zone are Cambro-Ordovician in age, comprising monotonous sequences of phyllites, quartzites and quartzo-phyllites with minor limestone layers (Zwart, 1979; 1986). Above these, the succession runs from black Silurian shales to Devonian slates and sandstones of variable thickness and finally to Carboniferous aged micaceous slates. This entire column of Palaeozoic sediments is strongly folded and unconformably overlain by younger rocks which are generally of Stephanian or younger age (Zwart, 1979; 1986).

Although the majority of these lower Palaeozoic rocks have been metamorphosed to low grade greenschist facies, in many areas a higher grade of metamorphism exists, with grades up to upper amphibolite facies. The rocks are generally mica-schists, but are occasionally seen to be migmatised in certain localities (Zwart, 1979; 1986).

Bodies of intrusive granodiorite of variable size are found scattered throughout the Axial Zone (Charlet and Dupuis, 1982). They are found in contact with rocks ranging in age from Cambro-Ordovician to Carboniferous age. Their distribution within the Axial Zone is shown in figure 1.5. Isotope studies have suggested a dominantly crustal origin (Zwart, 1979; Bickle et al., 1988). The timing of emplacement of these rocks, known locally as massifs, is subject to debate with some authors considering the intrusions to be synchronous with Hercynian deformation (Evans et al., 1997; 1998) whilst others consider intrusion to be a post-tectonic event (Zwart, 1979; 1986).

A number of the large massifs in the Pyrenees are composed of a homogeneous, leucocratic augen-gniess (Canigou, Hospitallet and Aston massifs). They represent deformed and metamorphosed granites of pre-Hercynian age (Zwart, 1986). Their exact timing of emplacement is also subject to debate, as they are covered by the majority of the Palaeozoic sequence. They may be intrusive relating to either the Cadomian orogeny (latest Pre-Cambrian to early Cambrian) or be intrusive granites of Cambro-Ordovician age (Zwart, 1986).
This described stratigraphy represents a stack of rocks up to 10 km thick that were all deformed during the Hercynian orogeny, during which the proto-Tethys and proto-Atlantic oceans were closed. The mode of deformation and sequence of events during the Hercynian is debated, with a variety of different structural models proposed to account for the observed structures. These range from rifting (e.g. Soula et al., 1986), to convergence (e.g. Zwart, 1979) or compression followed by extension (e.g. Vissers, 1992).

Despite the many different interpretations regarding the formation of these basement rocks, the massifs may be regarded as rootless plutons within a more ductile basement (Soula et al., 1986). In the context of this thesis the different mechanisms for the formation of the Hercynian structures is unimportant compared to that of the late stage exhumation of the massifs during the Alpine orogeny and their post-orogenic history.

A unique feature of the Pyrenees when compared to other Alpine aged orogens is that there is little deformation or metamorphism of the Hercynian basement, aside from at major fault zones (e.g. Mérens Fault, McCaig and Miller, 1986).

1.2.4.2 Mesozoic history

Following the Hercynian orogeny, sediments were deposited unconformably on the underlying lower Palaeozoic rocks. The events that took place during the upper Palaeozoic and Mesozoic are related to the position of the Iberian plate and its interaction with the European plate. The position of the Iberian plate at this time is given by figure 1.6a (late Carboniferous). The relative movement of Iberia was linked to the larger scale motions of the African and European plates, controlled by differential spreading rates and by the opening of the central Atlantic and the Bay of Biscay (Srivastava et al., 1990; Roest and Srivastava, 1991; García-Mondéjar, 1996). During the Mesozoic, extension was the predominant mechanism of deformation in the Iberian plate (Vergés et al., 2002). The Pyrenean rift was a ESE-WNW trending system and culminated with separation of Europe and Iberia and the opening of the Bay of Biscay (Pinet et al., 1987), with the Pyrenean rift connecting the Bay of Biscay in the west with the Tethys Ocean to the east (Vergés et al., 2002) (figure 1.6b). Initiation of rifting began in early Triassic (250 Ma), synchronous with the opening of the North Atlantic (Ziegler, 1990). Initially rifting resulted in the
development of a series of extensional, intra-continental basins, occurring along pre-existing Hercynian structures (Puigdefàbregas and Souquet, 1986). Following the development of these initial intra-continental basins, the transition to a fully marine setting occurred as a result of the continued extension between Europe and Iberia. Extension ceased in the mid Albian when sea floor spreading had begun in the Bay of Biscay (García-Mondéjar, 1996) and an anticlockwise rotation of the Iberian plate by \(\pm 22^\circ\pm 14^\circ\) had occurred (Moreau et al., 1997). In conjunction with the onset of seafloor spreading, motion along the plate boundary between Iberia and Europe changed from an extensional to a trans-tensional regime, causing strain to be accommodated in the development of pull-apart basins in the central Pyrenees. Trans-tensional motion was accommodated by a major strike-slip fault, the present day North-Pyrenean Fault. Activity on this strike-slip fault on a previously thinned crust led to magmatic events, with the emplacement of slices of upper mantle into the pull-apart basins (Fabriès et al., 1998). On the margins of the trans-tensional fault the post-extensional history is characterised by thermal relaxation of the lithosphere, leading to widening of the former rift basins (Vergés et al., 2002) (figure 1.6c).

1.2.4.3 Collisional history

Sea-floor spreading ceased in the Bay of Biscay during the mid-Santonian when the North Atlantic spreading centre extended further to the north. Following these events, Iberia collided with Europe marking the onset of the Pyrenean orogeny. Convergence began with the beginnings of subduction of the Iberian plate beneath the European plate (figure 1.6d).

Four main stages may be distinguished in the collisional evolution of the Pyrenees (Muñoz, 1992; Beaumont et al., 2000b). These are illustrated in the form of restored cross-sections in figure 1.7.

During the first stages of convergence, early Cretaceous extensional structures were inverted on both northern and southern sides of the orogen, whilst in the core of the chain, basement rocks began to become exposed (Beaumont et al., 2000b).

The second stage, Palaeocene, was a time of relatively slow plate convergence (Roest and Srivastava, 1991) and marks the time when foreland basins begin to develop on the margins
of the mountain belt, the Aquitaine basin to the north and the Ebro basin to the south (Puigdefàbregas and Souquet, 1986).

The third phase of convergence during the early to mid Eocene began with a rapid increase in the rate of thrusting. The foreland basins record widespread distributions of marine deposits (Puigdefàbregas and Souquet, 1986). In addition, during this time in the southern Pyrenees, successive thrust sheets developed in a piggy-back fashion with associated basins. These thrusts, the Boixols, Montsec and Sierras Marginales thrust sheets, detached on a décollement of weak Triassic evaporites (Puigdefàbregas and Souquet, 1986; Beaumont et al., 2000b). In the north some smaller thrust sheets did develop, but these were effectively pinned by the European plate.

During the fourth phase of convergence during the late Eocene and early Miocene, a change in deformational style occurred with deformation switching from thin skinned fold and thrust belts in the foreland to basement antiformal stack development in the core of the orogen. This antiformal stack is comprised of three main tectonics units. These units are named from top to bottom the Nogueres, Orri and Rialp units (Muñoz, 1992). During this time the first significant relief began to develop along the mountain belt and rapid exhumation occurred within the Axial Zone (Fitzgerald et al., 1999). In the northern Pyrenees the frontal thrust was forced to migrate 6 km into the foreland (Beaumont et al., 2000b). Continental sedimentation took place in the form of large conglomeratic deposits with sediment ponding occurring in intra-montane basins (Mellere, 1993). The last thrust movements in the southern foreland fold and thrust belt are given by the Sierras Marginales thrust at ~24.7 Ma (Meigs et al., 1996) and by apatite fission track constraints at the edge of the Axial Zone at 19.5±3.2 Ma (Sinclair et al., in press). This apatite fission track age is from the Barruera massif (see chapter 5 for location and discussion) and is the last documented period of active tectonics in the orogen.

Since ~20 Ma the Pyrenees have lain in a post-orogenic setting undergoing little further convergence, with tectonic activity effectively transferred in the Iberian plate to the Betic ranges to the south (Banks and Warburton, 1991; Vergés et al., 2002) (figure 1.6c). A total of 165 km of shortening has taken place in the central Pyrenees. This value is derived from numerical modelling and crustal balanced cross-sections (Muñoz 1992; Beaumont et al., 2000b).
Many of the observations and models described above have only been made possible by the large quantities of geophysical data that are available for the Pyrenean mountain belt and surrounding region. These include the ECORS deep seismic reflection line which crosses the central Pyrenees, from near Toulouse in the Aquitaine basin to Balaguer in the Ebro Basin (ECORS Pyrenean Team, 1988). This and a variety of other geophysical techniques such as gravity (e.g. Torné et al., 1989) and magnetotelluric surveys (e.g. Pous et al., 1995) have led to a greater understanding of the deep structure beneath the Pyrenean crust.

1.2.4.3.1 Syn-orogenic sedimentation

The well-preserved nature of the synorogenic deposits of the Pyrenees, particularly in the south, is one of the reasons that the Pyrenees has been extensively studied. Large bodies of coarse conglomerate are found overlying the southern fold and thrust belt, which reach a cumulative thickness of 3500m (Mellere, 1993). These sediments, known as the Collegats formation (Zwart, 1979), are interpreted as intra-continental basin deposits and in places such as the Sis conglomerate, as large long-lived palaeo-valleys acting as transfer zones of material away from the Axial Zone (Vincent and Elliot, 1997; Vincent, 2001). They represent a late Eocene to Oligocene alluvial fan and braided river complex and record the exhumation of the Axial Zone. In the northern Pyrenees, less well developed synorogenic deposits are recorded with only small exposures of conglomerates remaining today, for example the ‘Poudinges de Palassou’ (Vergés et al., 1995). In contrast, the excellent preservation in the south has led to theories of backfilling of the Ebro basin during deposition of these deposits and then later re-excavation during the mid to late Miocene (Coney et al., 1996). Controversy exists regarding this hypothesis (Gonzáles et al., 1997; Muñoz et al., 1997) as to whether an entire flank of a mountain belt could be draped in such a depth of conglomerates.
1.2.4.4 Foreland basins

1.2.4.4.1 Ebro Basin

The Ebro Basin is approximately triangular shaped and is bound to the north by the Pyrenees, to the south-west by the Iberian Ranges and to the south-east by the Catalan Coastal Ranges (figure 1.4). The formation of the Ebro Basin as a foreland system began in the Palaeocene by flexural subsidence related to the growth of the Pyrenees, the Catalan Coastal Range and the Iberian Range respectively (Garcia-Castellanos et al., 2003). Continued northward movement of the Iberian plate during the Palaeocene and early Eocene led to the emergence of the Pyrenees and Iberian ranges and closed off the connection to the Atlantic Ocean. This began a long period of endorheic (closed or internal) drainage and lacustrine sedimentation that persisted for millions of years until the mid-Miocene (Gaspar-Escribano et al., 2001; Garcia-Castellanos et al., 2003). During this time it is proposed that the synorogenic conglomerates described in section 1.2.4.3.1 backfilled the basin and draped the southern flanks of the Pyrenees (Coney et al., 1996).

During the late Miocene, the Ebro river breached the Catalan Coastal Ranges and widespread fluvial erosion occurred throughout the basin as it cut back through the filled foreland basin (Gaspar-Escribano et al., 2001; Garcia-Castellanos et al., 2003). The origin and timing of this major drainage change is poorly understood and relatively unconstrained with a variety of hypotheses, including river piracy and the Messinian Salinity Crisis suggested for the drop in base level (Garcia-Castellanos et al., 2003). The modern Ebro river drains the entire southern flank of the Pyrenees and reaches the Mediterranean Sea via a delta to the south of Barcelona. The modern system has been extensively dammed for hydroelectric purposes, but prior to this it was estimated that it carried $\sim 25 \times 10^6$ tonnes of sediment per year in 1880. Today, due to the extensive damming it only carries $0.12 \times 10^6$ tonnes, 2.5% of its former load (Evans and Arche, 2002; Guillen and Palanques, 1992).

1.2.4.4.2 Aquitaine Basin

In comparison to the Ebro Basin the Aquitaine Basin has undergone a considerably simpler history. The basin formed as a result of flexural loading by the Pyrenean mountains (Brunet,
Its development was strongly influenced by pre-existing Hercynian structural lineaments (Bourrouilh et al., 1995) leading to the inversion of mid Cretaceous pull-apart basins during the compressive deformation related to the formation of the Pyrenees. The sedimentary infill of the basin is composed of late Cretaceous to Eocene sandy-calcareous flysch sequences, becoming more siliciclastic during Eocene to Miocene times as the majority of the south and east of the basin became dominated by continental sedimentation (Borrouilh et al., 1995; Morris, 1999). In contrast to the Ebro Basin there is no evidence of basin closure to the ocean in the preserved deposits.

1.2.4.5 Post-collisional history

1.2.4.5.1 Neogene extension

From the mid-Miocene onwards a series of extensional intermontane basins developed in the eastern Pyrenees, for example, the Cerdanya basin (Cabrera et al., 1988; Roca 1996). These were associated with initially strike-slip and then later dip-slip normal faulting. These features have led to large flat regions in the modern landscape located between rugged mountains.

1.2.4.5.2 Modern tectonics

The Pyrenees are effectively a post-orogenic mountain belt. Although for the purposes of considering the post-orogenic evolution of the Pyrenees they may be considered as inactive tectonically, stresses are still present within the Iberian plate around the Pyrenees (Nicolas et al., 1990; Herraiz et al., 2000), (figure 1.6f) There is only a small element of compression as determined via earthquake focal mechanisms (Goula et al., 1999; Souriau and Pauchet, 1998).
1.3 Hypotheses and aims

The three controls of tectonics, climate and lithology on post-orogenic systems interact to shape the form of the orogen and the landscape. It is well accepted that during active orogenesis, tectonics is often the dominant force acting on the system (Willett et al., 2001; Burbank et al., 2003). Observations on modern mountain belts and results from both analogue and numerical modelling suggest that during active orogenesis an asymmetric form is developed (e.g. Stüwe, 1991; Koons, 1989; 1990; Beaumont et al., 1992; Willett et al., 1993; Kooi and Beaumont, 1996; Naylor, 2004), with a gentler pro-wedge and a steeper retro-wedge (figure 1.8b). This asymmetric form is related to the constant flux of material through the system and is stable over time. It has the effect of shifting the main drainage divide towards the retro-side of the mountain belt. It has been suggested that mountain belts in this type of form may reach a form of topographic steady state where the gross morphology of the system remains the same and simply downwears vertically as material is continuously fluxed through the system (Willett and Brandon, 2002). If, as in the case of many mountain belts where rivers are the principal agents of erosion, then in this topographic steady state, both rivers on the pro- and retro-sides of the orogen are at grade (Mackin, 1948). If either a detachment or a transport limited fluvial system is considered (Tucker and Whipple, 2002), then in a topographic steady state, differences in the values of area, slope and variables within the erosional efficiency factor across the principal drainage divide will balance one another out. This will result in equal amounts of downwearing on both the pro- and retro-sides of the orogen. This is true for a simple system where tectonics is the only driving force and smaller steeper catchments on the retro-side balance larger, more gently sloping catchments on the pro-side. This balance is illustrated in figure 1.8a where a simple cartoon illustrates the different parameters across the drainage divide.

If within this simple system, tectonics are shut off, what should happen to the topography subsequent to this tectonic cessation? The simplest hypothesis assumes that tectonics was the only control on the system and once stopped, and if the rivers on both pro- and retro-sides are at grade, then the mountain belt will simply downwear vertically at an exponentially decreasing rate. This would maintain the overall asymmetric form and keep the main drainage divide in a relatively constant location with respect to the mountain fronts. This null hypothesis is for a very simple system where tectonics are the only external forcing on the orogen and there is a homogeneous lithology and no climatic forcing. This hypothesis does make the assumption that the orogen is in a topographic steady state and that the orogen
has reached a level of maturity. However, even if this end-member case has not been reached, the presence of an asymmetric orogen and respective developed drainage systems would behave in a similar way. Therefore in summary the following null hypothesis can be stated:

**Null Hypothesis – following tectonic cessation, there will be steady topographic decay across the orogen.**

This hypothesis is for the simplest possible case and if heterogeneities in lithology are introduced or climatic forcing then alternative hypotheses may be followed:

**Alternative Hypotheses – In post-orogenic settings, lithology and climate will play an increasingly important role in controlling the macrogeomorphology and the resultant erosion history of the orogen.**

- The distribution of lithologies of differing resistance to erosion will control the position of the higher elevations.

- Present day asymmetry in precipitation has played an important role in reshaping the macrogeomorphic form of the Pyrenees.

- Pliocene climate change will have measurably rejuvenated the erosional record of the Pyrenees.

The primary aims of this thesis form a direct test of the above hypotheses:

- **To characterise the macrogeomorphic form of the Pyrenees at a catchment scale and to determine the relative controls on the form of the post-orogenic system.**

- **To characterise the timings and amounts of exhumation in the Pyrenees using low temperature thermochronology and the impact of this exhumation on the post-orogenic system.**
1.4 Structure of thesis

The thesis is divided into two principal sections:

The first section comprises chapters 2 and 3. Chapter 2 outlines the theory behind the use of digital elevation data and the techniques in which it is applied to the Pyrenees. Chapter 3 describes the results from the geomorphic study and considers them in light of the controls on the development of the form of the modern Pyrenees.

The second section comprises chapters 4 and 5. Chapter 4 describes the background to the use of both the apatite fission track and (U-Th)/He low temperature thermochronometers and the implications as to their use in mountainous regions. Chapter 5 describes the results from the thermochronometry study and presents thermal modelling results on the dataset. It also provides quantification on the amounts and timings of exhumation within the Pyrenees.

Finally, chapter 6 provides a synthesis of the above results and links the conclusions regarding the controls on the form of the Pyrenees with the timing and amount of exhumation derived from the thermochronology. The implications for these results are then considered, both on previous work on the Pyrenees and on post-orogenic systems in general.
Figure 1.1 - Main components of a simple mountain belt (After Morris, 1999)
Figure 1.2 - Stages of model development in the evolution of a simple mountain belt:

(1) Block Uplift bounded by step-up shear zones;
(2) Development of a low taper wedge over underthrusting plate;
(3) Development of a low taper wedge over overthrusting plate and verging in opposite direction;
S = Singularity
(After Willett et al., 1993)
Figure 1.3 - Location of Pyrenees within Western Europe
Figure 1.5 - Geological map of the Central Pyrenees, with principal drainage divide shown in black and political boundaries as dashed black line.
Figure 1.6 - Plate tectonic reconstruction of the Iberian Peninsula from the Upper Carboniferous to the present day. After Jabaloy et al., (2002). Present day emerged lands shown in green, oceanic crust in yellow. Main tectonic structures in black - dashes extensional; triangles compressional; and arrows indicating movement on transform structures. Active spreading axis shown by thick black line and inactive spreading axis by thick dotted black line.
Figure 1.6 (continued)
Figure 1.7(a) Balanced, partially restored and fully restored cross-sections along the ECORS profile (After Beaumont et al., 2000b).
Figure 1.7(b) Balanced, partially restored and fully restored cross-sections along the ECORS profile (After Beaumont et al., 2000b).
Figure 1.8 - Cartoon and cross-section through a mountain belt in topographic steady state. Rivers on both pro- and retro-sides are at grade.