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Fluvial evolution of the Garonne River, France: integrating field data with numerical simulations

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FLUVIAL EVOLUTION OF THE GARONNE RIVER, FRANCE: INTEGRATING FIELD DATA WITH NUMERICAL SIMULATIONS

A Thesis

Submitted to the Graduate Faculty of the Louisiana State University and Agricultural and Mechanical College in partial fulfillment of the requirements for the degree of Master of Science in The Department of Geology & Geophysics

by

Robin Rene Lancaster
B.S., Colorado State University, 2003
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ABSTRACT

The Garonne River of southwestern France presents a unique opportunity to study the controls on long-term incision and terrace formation within a large-scale fluvial system. The Garonne heads in the Pyrenees, flows through the Aquitaine Basin, and discharges into Atlantic Ocean via the Gironde Estuary/Bay of Biscay. From field data, three terrace complexes were identified and traced for >251 km from the base of the Pyrenees to the onset of tidal influences. Each complex is separated from adjacent complexes by scarps >10 m and represent 10s to 100s kyrs during which time the river occupied a finite elevational range within its valley, and lateral migration and cutting of straths dominated over valley incision. Using incision rates determined from other European studies, the 3rd terrace complex likely formed 500-250 kyr BP, the 2nd terrace complex formed 200-100 kyr BP, and the 1st terrace complex formed 100-50 kyr BP. Without direct chronological control, it is not possible to infer correlation between periods of deposition or bedrock incision and specific climatic conditions. Following previous work, a numerical model was developed to test the dependence exerted by slope and discharge on incision. Hypotheses tested include detachment-limited, transport-limited, and total stream power relationships, as well as a null hypothesis where incision is independent of discharge and slope. Each model has a specific range of exponents for discharge and slope ($m$ and $n$ values, respectively) within the overall incision equation. Error and fit for each model formulation were evaluated using statistical tests. Results from the detachment-limited and transport-limited models were all unacceptable, whereas the total stream power model produced three acceptable combinations of $m$ and $n$. However, the best-fit result for this system is the null hypothesis. It is therefore concluded that the models tested in this study do not describe the hydraulics behind incision in this river, or valley incision is independent of slope and discharge. Future work
should focus on refinement of models so as to test for differential uplift and variations in lithologic controls, and development of independent geochronological control so as to evaluate causal mechanisms.
INTRODUCTION

River systems are the primary agents for erosion of the landscape, and major river valleys are the primary conduits for transport of sediments to ocean basins. A number of recent studies have shown that time-averaged rates of erosion and sediment delivery to the oceans have remained relatively constant when considered over time scales of $10^6$ yrs or more (see Phillips, 2003). However, it has long been known that over shorter time scales, erosion of landscapes, river valleys in particular, is not a continuous process when measured in terms of rates, but instead consists of periods of bedrock valley incision punctuated by periods of lateral migration of channels and construction of floodplains (Merritts et al., 1994; Blum and Törnqvist, 2000). Renewed valley incision then leaves this former floodplain behind in the landscape as a terrace. As a result, flights of terraces that line major river valleys represent a series of timelines in the landscape, and a record of both the long-term trend and the discontinuous nature of fluvial landscape evolution.

This thesis examines fluvial system evolution by focusing on terraces of the Garonne River of southwestern France. The Garonne heads in the moderately glaciated Pyrenees, and its upper reaches demonstrate bedrock erosion typical of mountainous streams. As the river travels towards its termination in the Gironde Estuary and the Bay of Biscay/Atlantic Ocean, the upland incisional regime changes to mixed depositional and erosional regime typical of a coastal plain incised valley system proximal to a marine basin. With such diversity, the Garonne presents a unique opportunity to study how a large fluvial system evolves over time, and responds to the interaction between tectonic or isostatic uplift and climate change in upstream sediment source regions, and sea-level change in the lowermost reaches.
Specific goals of this thesis are stated as follows:

1. Construct cross-sectional and longitudinal profiles of the Garonne terraces from published geologic maps, topographic maps, digital elevation models, and field reconnaissance, and verification of map units.

2. Develop a series of numerical models to calculate and describe long profile changes in the Garonne Valley then compare model results to empirical records. In particular, this includes (a) examination of the interactions between climate and sea level controls, using mapped terraces as reference points, and (b) estimates of the age of three terrace complexes, based on empirically-derived erosion laws and comparison with similar systems within Europe.

3. Determine, from numerical model results, what drives large-scale erosion based on three hypotheses:
   
   1) Patterns of valley incision are best described by unit stream power or shear-stress parameters
   
   2) Patterns of valley incision are best described by transport-limited parameters
   
   3) Patterns of valley incision are best described by total stream power parameters

In addition to the above specific goals, this thesis is part of a larger research program that includes studies of the Loire and Rhone Rivers, also in France. Sediment samples from the Garonne valley were collected for optically-stimulated luminescence dating, and will be processed at some time in the future. Results will be included in the database from this larger research program, but are not part of this thesis.
**GENERAL SETTING**

The Garonne River flows 525 km north-northwest from its headwaters within the Aran Valley, Spain, in the high Pyrenees, to the Gironde Estuary, France, which discharges the Atlantic Ocean via the Bay of Biscay (Figure 1). The Garonne is the largest river in southern France, drains ~57 000 km$^2$ of the Pyrenees, Massif Central, and Aquitaine Basin physiographic regions, and descends 1900 m in elevation from source to mouth. The majority of the Garonne is located within France, its headwaters only ~50 km inside the Spanish border. Major cities along the Garonne main stem include Toulouse, Agen, and Bordeaux, where the Garonne joins the Dordogne to form the 75 km-long Gironde Estuary (UNEP/DEWA, 2004).

**Geologic Setting**

Southwestern France owes its topographic complexity to its tectonic history. Like most of Europe, this region is underlain by Precambrian igneous and metamorphic basement rocks, which are now mostly exposed in a series of massifs that represent the cores of Devonian through Carboniferous (Hercynian) orogenic belts. These include the Massif Central of southcentral France and the American Massif of northwestern France (Ager, 1980). During the Mesozoic and Cenozoic, the southern and western European continental margins accumulated a succession of sedimentary rocks that lap onto the major massifs. Early Cenozoic convergence between the Eurasian and African plates then led to closure of the Tethys Sea along the southern margins of Eurasia, formation of major collisional mountain belts in southern Europe, such as the Alps and Pyrenees, and formation of a series of associated sedimentary basins, such as the Aquitaine (Ager, 1980; Embleton, 1984). Late Cenozoic erosion of the massifs and mountain belts resulted in thick sediment accumulation within basins such as the Aquitaine.

The Pyrenees and Massif Central are the two primary topographic elements that control
Figure 1. Location map. The dashed black line denotes the border of the Garonne River drainage basin.
of the Garonne heads in the igneous and metamorphic core of the Pyrenees, which extends east
to west along the border between France and Spain: the upper Garonne, as defined here, lies
wholly within the Pyrenees, and flows through a bedrock channel typical of mountainous
regions. The upper Garonne was glaciated numerous times in the Pleistocene (see Calvet, 2004,
for a review), hence the valley contains a series of moraines and glacial outwash terraces.

The middle Garonne, as defined here, begins at the Pyrenees front, then traverses a
moderately low-relief landscape dominated by Cenozoic sedimentary rocks of the Aquitaine
Basin. The Aquitaine is an asymmetric retro-foreland basin, a result of the influence of loading
in the Pyrenees mountains and piedmont (Verges et al., 2002) (Figure 2), with Precambrian to
late Paleozoic basement at depths as great as 7000 m near Bordeaux (Ager, 1980). Major folds
within the basin trend north-northwest, perpendicular to the Pyrenees, and can be attributed to
four major folding episodes within the basin, the most important of which occurred during the
Eocene (Ager, 1980). The middle Garonne flows along such a fold, which marks the boundary
between mostly calcareous sedimentary rocks to the north and unconsolidated Tertiary sands
and gravels in the south (Ager, 1980). The middle Garonne channel would be classified as
mixed bedrock-alluvial stream (sensu Howard et al., 1994; Howard, 1995), and the valley
contains a classic flight of terraces that represent episodic bedrock valley deepening punctuated
by lateral migration of deposition of coarse gravels and sands.

For the purpose of this thesis, the middle Garonne is separated into three sections, 1) the
upper-middle section which extends from the Pyrenees front to the confluence of the Ariège
and Garonne south of Toulouse at Portet sur Garonne, 2) the central section which includes the
Garonne from Portet sur Garonne to Agen, encompassing the confluence with the Tarn River,
and 3) the lower-middle section which extends from Agen to Castets, including the confluence
with the Lot River. This thesis focuses primarily on these three sections.

The lower Garonne, as defined here, begins at Castets, where tidal influences are first felt, and extends to the confluence with the Dordogne at Bordeaux and the Gironde Estuary. The Gironde Estuary is macrotidal, with a tide range at Bordeaux of 2.5 (neap tide) to 5 m (Spring tide) (Allen and Posamentier, 1993). This part of the valley also lies within the Aquitaine Basin, and contains a flight of well-defined terraces.

The Massif Central forms a divide between southern France’s two other major fluvial systems, the Rhone, which discharges to the Mediterranean, and the Loire, which discharges to the Atlantic (Embleton, 1984). Major tributaries to the Garonne, including the Lot and Tarn Rivers, also head in the Massif Central, then flow to the west before joining the Garonne.

**Climate and Hydrology**

The climate of southwestern France closely reflects general circulation of the atmosphere, the corresponding location of semi-permanent high and low pressure centers, and the influence of the Pyrenees (Benito, 2003), and is intermediate between the Mediterranean climate regime farther south and east, and the maritime or oceanic regime farther north and west (Kendrew, 1957; Plaut and Simonnet, 2001). The upper Garonne (within the Pyrenees) is commonly below freezing during the winter months, with mild summer temperatures. Farther downstream, within the Aquitaine Basin, winter temperatures are mild and rarely below freezing, whereas summers are hot. Mean January temperatures at Andorra la Vella, Andorra, in the Pyrenees are 0 °C, with values of 5 °C at Toulouse and Bordeaux. Mean July temperatures at Andorra la Vella, Andorra, in the Pyrenees are 16 °C, with values of 22 °C and 20 °C at Toulouse and Bordeaux, respectively (EuroWeather, 2005) (Figure 3).
Figure 2. Geologic setting. Generalized geologic map of France. After Kirkaldy, 1967.
Mean annual precipitation in the Garonne watershed is variable and reflects the influence of the Mediterranean and maritime climate regimes with high precipitation in the upper and lower reaches, and lower precipitation in the middle reaches of the Aquitaine basin. The high Pyrenees receive substantially more precipitation than the southern portion of the Aquitaine basin, with values of 70.8 mm/yr at Andorra la Vella, Andorra, 56.3 mm/yr at Toulouse, and 80.6 mm/yr at Bordeaux (EuroWeather, 2005). Seasonality of precipitation is significant with ranges of 50 mm in the Pyrenees at Andorra la Vella, Andorra, and 32 mm and 36 mm at Toulouse and Bordeaux, respectively (Figure 4).

Mean annual discharge of the Garonne ($m^3/s$) is ~200, ~425, and ~600, at Toulouse, Agen, and Langon, respectively (S.M.E.P.A.G., 1989). The highest average monthly discharge occurs during May for the middle and upper reaches, averaging 405 $m^3/s$ and 707 $m^3/s$ at Toulouse and Agen, respectively. Discharge peaks in the lower Garonne in April with an average of 1086 $m^3/s$ at Langon (Pardé, 1935). Floods for the upper and middle sections are linked to both rain and snowmelt, while the lower section floods primarily as the result of heavy rainfall. Centennial floods discharge 4700 $m^3/s$, 6800 $m^3/s$, and 7700 $m^3/s$, at Toulouse, Agen, and Langon, respectively, and five-year floods discharge 2450 $m^3/s$, 4120 $m^3/s$, and 4700 $m^3/s$, at Toulouse, Agen, and Langon, respectively (S.M.E.P.A.G., 1989)(Figure 5).

Past climates of Southern France are also well understood due to the continuous and temporally significant climatic records from a series of crater (maar) lakes located within the Velay mountains of the Massif Central and bogs located north of Lyon. These include Lac du Bouchet (e.g. Thouveny et al., 1994; Reille and de Beaulieu, 1995; Tzedakis et al., 1997; Williams et al., 1998; Roger et al., 1999; Stockhausen and Thouveny, 1999), Praclaus crater (e.g. Reille and de Beaulieu, 1995; Tzedakis et al., 1997; Roger et al., 1999), Les Eschets bog
Figure 3. Average monthly temperature at select locations. Average monthly temperature as measured at Bordeaux (lower Garonne), Toulouse (middle-Garonne), and Andorra La Vella (upper Garonne).
Figure 4. Average monthly precipitation at select locations. Average monthly precipitation as measured at Bordeaux (lower Garonne), Toulouse (middle-Garonne), and Andorra La Vella (upper Garonne).
Figure 5. Hydrology information at select locations. Hydrology of the Garonne River system as measured at gauging stations in Toulouse (upper-middle), Agen (central), and Langon (lower-middle).
(e.g. Guiot et al., 1989), Lac St-Front (e.g Thouveny et al., 1994, and Stockhausen and Thouveny, 1999) and La Grande Pile bog (e.g. Guiot et al., 1989). These lakes and bogs provide a continuous record nearly 500,000 years long. Using data as varied as magnetic susceptibility and palynology, paleoclimate signatures have been reconstructed and correlated with global and hemispheric signals, showing a strong correlation with Greenland ice core records (e.g. Thouveny et al., 1994; Williams et al., 1998; Stockhausen and Thouveny, 1999; Klotz et al., 2004) and with the SPECMAP composite marine oxygen isotope curve developed from the Atlantic, Indian, and Pacific ocean basins (Roger et al., 1999)(Figure 6).

At a more detailed level, maar lake and bog records mentioned above have been correlated using a tephra erupted ca. 285-290 kyr BP, during marine isotope stage (MIS) 9a (e.g. Imbrie et al., 1984; Reille and de Beaulieu, 1995; Williams et al., 1998). Both pollen and magnetostratigraphic data from French lakes and bogs show at least six major interglacial periods during the last 400 kyr, excluding the current Holocene Interglacial (Reille and de Beaulieu, 1995; Roger et al., 1999; Raynaud et al., 2005). These include the Praclaux interglacial (MIS 11b, ~400 kyrs BP), the Jagonas complex interglacial (MIS 11, ~400 kyrs BP), the Landos interglacial (MIS 9c, 315-300 kyrs BP), the Amargiers interglacial (MIS 9a, 280-290 kyrs BP), the le Bouchet complex interglacial, composed of Bouchet I,II and III episodes (MIS 7a-7c, 240-190 kyrs BP), and the Ribains (referred to as Eemain in Northern Europe) interglacial (MIS 5e, 128-115 kyrs BP) (Reille and de Beaulieu, 1995; Roger et al., 1999). Although not the focus of this thesis, the presence of detailed paleoclimate records such as these will ultimately facilitate understanding of the role of climate change in fluvial system evolution.
Figure 6. Climate records for the last 350 kyr BP. (A). The average Lac du Bouchet susceptibility is composed of four parallel records. (B). Arboreal/non-arboreal pollen ratio. (C). High resolution oxygen isotope records from the North Atlantic Ocean. (D). High resolution oxygen isotope records from the Pacific Ocean. (E). High resolution oxygen isotope records from the Indian Ocean and identified isotope stages 1 to 9 from the last 300 kyrs. After Roger et al., 1999.
REVIEW OF LITERATURE

Historical Perspectives and Contemporary Studies of Fluvial Landscape Evolution

As noted above, fluvial systems are the primary agent for erosion of the landscape and transport of sediments to sedimentary basins (Bridge, 2003). Over long time scales, rates of erosion of the landscape and deposition in sedimentary basins remain relatively steady, and reflect mostly tectonic controls. Over shorter time scales, rates are more variable and likely reflect climatic controls and climate change (Blum and Tornqvist, 2000; Bridgeland, 2000; Bridge, 2003). Long-term fluvial landscape evolution, and the shorter-term response of fluvial systems to climate change is therefore an important topic. However, large river systems like the Garonne are poorly understood with respect to long-term evolution.

Flights of terraces are widely recognized to represent a river’s long profile evolution through time in response to external controls, such as climate change, sea level change, and tectonic activity, and have been a focus of study for more than 100 years. As defined by Merrits et al. (1994), a terrace is an abandoned surface that was once the active river floodplain, and can be subdivided into the terrace tread, scarp and strath (Figure 7): the tread is defined as the surface (generally depositional) that represents the formerly active floodplain, although some terraces include eolian and other sediment covers as well (see Blum et al., 2000), whereas the scarp is defined as the slope that connects the tread to a lower surface, such as the tread of a younger terrace or an active floodplain. Scarps therefore represent floodplain abandonment and incision to a lower level within the valley. Terraces are, in turn underlain by straths, which is the surface cut into bedrock, and which is normally overlain by fluvial deposits that represent lateral migration or aggradation by the river channel. Accordingly, straths represent the maximum depth of bedrock valley incision at a particular time.
Figure 7. Terrace definition diagram. Strath, tread and scarp surfaces are labeled. The upper scarp is less steepened because eroded materials from the top of the scarp have been deposited at the toe. Initially, the upper scarp resembled the lower scarp.
Since that time, a number of workers have suggested that both upstream and
downstream controls may influence terrace generation, though at different positions along the
stream profile (e.g. Blum and Valastro, 1994; Merritts et al., 1994; Törnqvist, 1998; Bridgland,
2000; Pazzaglia and Brandon, 2001). Current work chronicles the utility of terraces to develop
stratigraphic frameworks in major fluvial systems of Western Europe, including the Rhine-
Meuse (van den Berg, 1996; Törnqvist, 1998) the Loire (Straffin et al., 1999; Blum and
Straffin, 2001; Straffin and Blum, 2002), the Rhone (Mandier, 1988), the Ebro (Fuller et al.,
1998), the Thames (Maddy et al., 2001) and others. It is safe to say that the use of terraces as an
interpretive aid for studies of landscape evolution and response to climate change and other
system controls has become more prevalent in recent years (e.g. Blum et al., 2000; Bridgland,
2000; Vandenberghe, 2003). In part, this reflects the development of new geochronological
techniques, such as optically stimulated luminescence (OSL), and cosmogenic nuclides that
provide for development of chronological frameworks in systems that previously had
inadequate age constraints (Mejdahl and Funder, 1994; Wintle, 1997; van Heteren et al., 2000;
Schaller et al., 2002; Rittenour et al., 2003; and many others).

Terraces also have received considerable attention in the context of development of
landscape evolution models. Hack (1957) first used long profiles of fluvial terraces to interpret
the relative importance of relief and rock strength on rates of channel incision, and suggested
that river long profiles remain relatively constant over long periods of time. A number of recent
workers have built on these concepts, implicitly recognizing that terraces represent a series of
timelines in the erosional landscape (e.g. Merritts et al., 1994) and, in theory, can provide
important keys to understanding rates of uplift and bedrock incision, as is demonstrated by
Pazzaglia and Brandon (2001), in their study of the Clearwater River in Washington State, and
by Schaller et al. (2002) in their study of erosion rates in Middle European rivers.

In this context, numerical modeling is a relatively new and rapidly evolving set of techniques for qualitative and quantitative simulation of earth-surface processes (Tucker and Slingerland, 1994). Although there remain numerous assumptions and problems (see Hovius, 2000; Blum and Tornqvist, 2000; Bogaart and van Balen, 2000; Hancock and Anderson, 2002; Tomkin et al., 2003), numerical modeling provides a way to use process-based relationships to visualize and manipulate various controls in ways that would otherwise be impossible. Terraces, because they represent a series of timelines in the erosional landscape, provide an opportunity to test and calibrate numerical landscape evolution models. Two approaches are used to create such models. One uses physically-based rules in an effort to test fundamental principles behind fluvial evolution (e.g. Tucker and Slingerland, 1994; Hancock and Anderson, 2002; Tomkin et al., 2003). The other uses process-based rules and relationships to test specific parameters (Veldkamp, 1992). Both models have advantages, but physics-based models make fewer, or at least more fundamental assumptions, and therefore tend to be less qualitative.

**Previous Work on the Garonne River**

Relatively few studies have focused on terraces within the Garonne valley, or correlations between terrace formation and specific climatic or sea level controls. Early work by Hubschman (1971, 1972, 1973, 1975, 1984) and Hubschman and Le Ribault (1972) suggested that Garonne terraces, mostly those upstream from Toulouse, represent the glacial-interglacial cycles of the Pleistocene (Figure 8). The upper-middle Garonne was studied in depth by Hubschman (1975), who correlated the floodplain, and 1\(^{st}\), 2\(^{nd}\), and 3\(^{rd}\) terrace to the Würm (MIS 2-4), Riss (MIS 6), Mindél (pre-Riss), and Günz (pre-Riss) glacial periods, respectively as classically defined by the Penck and Bruckner (1909) model. More recent
research, conducted within the context of our present understanding of glacial and climate system history (e.g. the SPECMAP and ice core records), has not been undertaken within the Garonne system. Geologic mapping by the Bureau de Recherches Géologiques et Minières (BRGM, French Geological Survey) identified a range of terraces in the Garonne Valley, though correlations and age relations between maps in different parts of the valley remain uncertain and in some cases inconsistent. New techniques such as numerical experiments and OSL dating allow for a more precise determination of terrace ages and commentary on the timing of their formation.

In contrast to the Garonne valley, considerable research has been conducted on the morphology and dynamics of the Gironde Estuary (see Maillet et al., 2000; Steiger et al., 2001; Schäfer et al., 2002). Stratigraphic evolution of the Gironde has also been examined in some detail, and is the basis for incised valley models in macrotidal esturine settings (Figure 9)(e.g. Allen and Posamentier, 1993; Allen and Posamentier, 1994). Formed as a result of sea level fall during the Holocene, deposits accumulated within the Gironde Estuary represent low-stand, transgressive and high-stand systems tracts, of which transgressive valley fills are most prevalent. During transgression, the incised valley is flooded to form an estuary. This is followed by landward migration of the estuary mouth, forming a tidal ravinement surface then a wave ravinement surface after which the sequence is capped by a maximum flooding surface (Allen and Posamentier, 1993; Allen and Posamentier, 1994).

The offshore extension of the Gironde has been studied with seismic data, and limited boreholes (Pinet et al., 1987; Lesueur and Tastet, 1994; Lesueur et al., 2002). In general, these studies show mid-shelf mud fields deposited within the last 2 kyrs, overlying fine-medium sand and pebbly shelf materials deposited 5 kyrs BP on a thinning Aquitaine shelf, the result of
extension of the Parentis sedimentary basin, a sub-basin within the Aquitaine basin (Bourrouilh et al., 1995). Mud accumulation is the result of both episodic flooding of the estuary and continuous settling, while sand deposition results from storm reworking of nearby sandy shelf deposits.
Figure 8. Previous mapping of the upper-middle Garonne. Map of upper-middle Garonne terraces and sediments from Hubschman’s 1975 study of the Garonne. The high terrace corresponds to the 3rd terrace complex from this study, the middle terrace to the 2nd terrace complex, the low terrace to the 1st terrace complex, and the floodplain to the floodplain complex. After Hubschman, 1975.
Figure 9. Estuarine model developed from the Gironde Estuary. This model shows the stratigraphic relationship between transgressive aggradational estuarine coastal-plain deposits and the alluvial plain. The relationship depicted above exists synchronous to landward bayline migration, such as would occur during sea-level rise. After Allen and Posamentier, 1993 as modified by M. Blum.
METHODS

Field Techniques

Three terraces plus the modern floodplain were identified from BRGM maps and traced for ~331 km from the base of the Pyrenees at the village of Cazeres, to the mouth of the Gironde Estuary northwest of Bordeaux. Preliminary long profiles of terrace surfaces were constructed in the field from 1:100,000 and 1:25,000 topographic maps published by the Institute Geographique Nationale (IGN), and 1:50,000 geologic maps published by BRGM.

In contrast to other terrace studies, this thesis does not contain detailed sedimentological descriptions of the floodplain and three terrace complexes. This is due in part to poor exposures of said complexes and because such information is not the focus of this thesis.

Numerical Techniques

Numerous recent studies have investigated the role of various hydraulic parameters such as stream power or shear-stress (see Singh, 2003 for a review) in fluvial incision models (e.g. Grant, 1997; Whipple and Tucker, 1999; Lisle et al, 2000; Tucker and Whipple, 2002; Baldwin et al., 2003; Synder et al., 2003; Tomkin et al., 2003; Finnegan et al., 2005). These studies have focused mainly on small bedrock catchments, but there is increasing interest in using such equations to model alluvial streams as well (Howard et al., 1994; Sklar and Dietrich, 1998; Sklar and Dietrich, 2001), especially as sediment supply rates and the erodibility of “covered” beds is more understood (Sklar and Dietrich, 2001). Most of the models currently in use employ some derivation of a power law that relates incision rate, drainage area (or discharge) and channel gradient (Tucker and Whipple, 2002). Unfortunately, tests of these theoretical models on real-world rivers have shown that they are often unable to describe natural processes (e.g. Stock and Montgomery, 1999; Tomkin et al., 2003). Theories proposed
for these failures include improper scaling (Finnegan et al., 2005), imprecise measurements of sediment supply, grain size, and rock strength (Sklar and Dietrich, 2001) and higher than average localized stresses (Lisle et al., 2000). One of the primary goals of this research is to test these models on the Garonne River.

In its upper reaches, the Garonne River is a bedrock river incising into the Pyrenees. The central and lower portions of the Garonne, in contrast, are alluvial, flowing through and transporting the Tertiary sediments that fill the Aquitaine basin. However, it must be considered that terraces, by definition, are the result of incision into bedrock, which may or may not have been followed by deposition (Bull, 1991; Merritts et al., 1994). The terraces of the Garonne therefore suggest that the modeled stretch is now alluvial to mixed bedrock-alluvial, but it must have been a bedrock channel at the time of terrace generation. Therefore, it is appropriate to test relationships that describe both bedrock and alluvial rivers, as the timing of the transition between the two is unknown, and is generally a poorly understood concept thought to be related to transport capacity limitations (Whipple and Tucker, 2002). It should be noted that this is the first time the terraces of this river have been modeled and the first time any model has been used on a basin of this scale (55,000 km²).

Physically-based empirical models are used to examine large-scale river-wide changes over time, as expressed by changes in bed elevations and long profiles. To be precise, rivers should be modeled with non-linear equations such as the Navier-Stokes equation, which describes three-dimensional non-uniform unsteady flow (Bogaart and van Balen, 2000). However, assumptions of steady uniform flow (Vining 1998), constant stream power, uniform lithology, sediment density and grain size, constant sediment discharge, and constant water discharge are typically used to facilitate computations because of the long time scales involved.
in long-term incision rates. A subset of these assumptions are used in this study as well.

Following previous work (e.g. Stock and Montgomery, 1999; Tomkin et al., 2003; van der Beek and Bishop, 2003) a numerical model for changes in bed elevation through time along the Garonne River was developed to test Willgoose et al.’s (1991) sediment transport equation, which relates sediment discharge to changes in water discharge and slope using the following power equation:

\[ q_s = \beta q^{m'} S^n \]  

Where:

- \( q_s \) = sediment discharge (m\(^2\)s\(^{-1}\))
- \( \beta \) = multiplicative constant
- \( q \) = water discharge (m\(^2\)s\(^{-1}\))
- \( S \) = slope in steepest downhill direction (m/m)
- \( m' \) = power of \( q \)
- \( n \) = power of \( S \)

This equation is a derivation of the Einstein-Brown equation, which relates sediment transport to shear stress. To obtain this form of the equation, sediment is considered to be homogenous throughout the basin and small approximations are made about the representative grain size and specific gravity of sediments, and the density of water in the catchment area.

For this study, bed elevation changes over time (incision or aggradation rate) were equated to sediment discharge:

\[ \frac{dz}{dt} = K q^m S^n \]  

Where:

- \( \frac{dz}{dt} \) = change in bed elevation over time (m/s)
- \( K \) = multiplicative constant (m\(^2\))
- \( q \) = water discharge (mean annual discharge: m\(^3\)s\(^{-1}\))
\[ S = \text{channel slope (m/m)} \]

\[ m = \text{power of } q \]

\[ n = \text{power of } S \]

Bed elevation changes over time, \( dz/dt \), are equivalent to \( q_s \) per unit length, as follows:

\[ dz/dt = q_s/l \]  \hspace{2cm} (3)

Where:

\[ dz/dt = \text{change in bed elevation over time (m/s)} \]

\[ q_s = \text{sediment discharge (m}^2\text{s}^{-1}) \]

\[ l = \text{length of stream (m)} \]

This relationship is maintained when averaged over sufficiently long time scales (Willgoose et al., 1991), which for this study are assumed to be time scales of \( >10^4 \) yrs. In addition, \( m' \) in the Willgoose et al. (1991) equation is equivalent to \( m \) used in this study, such that \( m' = 2m + 1 \). The boundary limits of the transport-limited and unit stream-power equations from the literature (i.e. Whipple and Tucker, 1999; Whipple et al., 2000; Tomkin et al., 2003) are in \( m' \) form, but for simplicity \( m \) is used henceforth in all hypotheses and results. This simulation tests for values of \( 2 \geq n \geq 0 \) and \( 1 \geq m \geq -0.5 \) (\( 3 \geq m' \geq 0 \)), which are used to relate basin hydrology, hydraulic geometry, and erosion processes (Whipple and Tucker, 1999). The range chosen for this study allows for the comparison of commonly tested models with other scenarios, such as a situation when \( m > n \).

For this study, the following three hypotheses were tested to examine the dependence of the river system on changes in discharge, \( q \), and slope, \( S \), as follows:

(1) Patterns of valley incision follow a shear-stress or unit stream power relationship: This hypothesis predicts a system that is more sensitive to slope than discharge where rates of erosion are non-uniform. This power-relationship is designed to test detachment-
limited parameters and would apply to the Garonne only if the Garonne was behaving as a bedrock stream at the time of strath-cutting. For incision to occur, the threshold of basal shear stress must be overcome. This model analyzes a range of empirically-derived units to describe this relationship. If this relationship were true for this system, $m$ and $n$ would behave such $2m = n$ over the range $2.5 \geq n \geq 0.66$ (Howard et al., 1994; Tucker and Slingerland, 1997; Whipple et al., 2000; Whipple and Tucker, 2002). These values for $n$ are theoretical representations of the processes of plucking (low values) and suspended-load abrasion (high values) (Snyder et al., 2003).

(2) Patterns of valley incision follow a transport-limited relationship: Rates of erosion are non-uniform and long profile evolution behaves according to transport-limited rules, where the system is more sensitive to slope than discharge. The transport-limited relationship describes incision as the result of the ability of the stream to transport sediment. This is different than the detachment-limited relationship, as described above, in that sediment supply is tested within the parameters set forth in this model. In this relationship, incision occurs when the amount of sediment being transported falls below the transport capacity of the river. If the transport capacity is exceeded (sediment volumes are very high), deposition occurs. This hypothesis predicts values of $0 \geq m \geq -0.5$ and $2 \geq n \geq 1$ (Tucker and Slingerland, 1997; Whipple et al., 2000; Whipple and Tucker, 2002; Tomkin et al., 2003).

(3) Patterns of valley incision follow a total stream power relationship: Long profile evolution occurs through uniform rates of erosion throughout the studied reach. This relationship is often referred to as stream power per unit channel length, and predicts that $m=n=1$ (Seidl and Dietrich, 1992; Whipple and Tucker, 2002), or at the very least
Total stream power incises by virtue of a stream’s velocity, slope and the density of the sediment-water fluid (Bull, 1991). As with the shear-stress/unit stream power model, this relationship tests the ability of a stream to detach sediment from the bed. However, it also tests the transport capacity of the river, making it similar to the transport-limited model. This model places equal importance on both discharge and slope by assigning equal values to their exponents of $m$ and $n$.

The null hypothesis for this study can be defined as a situation in which $m = n = 0$. This implies that the model is not able to describe incision in this system, or that incision of the Garonne into its bedrock is independent of changes in slope and discharge. If the latter were true, it would also imply that physical weathering is a more important process than previously thought (Whipple et al., 2000).

The equations used here and in the literature to test various models, such as transport- or unit stream power limited parameters, are based on the physically-derived parameters of shear stress ($\tau$), total stream power ($\Omega$), and unit power ($\omega$). Shear stress is simply the force per unit area exerted on the bed by the moving fluid in the direction parallel to flow, as defined below in equation (4). Total stream power is defined as the ability of a river to perform work, in this case transport bedload materials, and is equally dependent on discharge and slope (equation 5 below); as either increase, so does the capacity of the stream to transport bedload and suspended load sediments. Within a system that behaves in this manner, the downstream increase in discharge is matched by the downstream decrease in slope, resulting in uniform incision. Incision occurs when the total stream power is larger than flow resistances. Conversely, aggradation occurs when flow resistance is greater than total stream power (Bull, 1991). As previously noted, parameters such as $\gamma$ are assumed to be constant and are figured
into the K of equation (2) used to model the fluvial development of this river system. Unit stream power is simply total power divided by channel width. Shear stress ($\tau$), total stream power ($\Omega$), and unit power ($\omega$) are defined by:

$$\tau = \gamma dS \quad (4)$$

$$\Omega = \gamma QS \quad (5)$$

$$\omega = \Omega/w = \gamma QS/w = \gamma dSU = \tau U$$

Where:

- $\tau$ = shear stress
- $\Omega$ = total stream power
- $\omega$ = unit stream power
- $\gamma$ = specific weight of sediment-water fluid
- $Q$ = stream discharge
- $S$ = channel slope
- $w$ = streambed width
- $U$ = mean flow velocity

Shear stress and unit stream power are functionally equivalent in calculations, as depth and velocity are assumed to be constant for the stretch of river modeled, and width is assumed to vary proportionally with discharge. A stream that is described by a shear stress or unit stream power equation therefore assumes that long profile evolution is more dependent on slope than discharge. Thus, in incision equation 2, $m$ and $n$ would behave such that $2m = n$ over the range $2.5 \geq n \geq 0.66$ (van der Beek and Bishop, 2003).

Transport limitations restrict incision rate by limiting the ability of the stream to transport eroded materials rather than detach them (Whipple and Tucker, 2002; Tomkin et al., 2003). A stream adhering to transport-limited rules is more dependent on slope than discharge,
like the shear stress or unit stream power case, but differs in that $m$ and $n$ have a specific range of acceptable values ($0 \leq m \leq 0.5$ and $2 \leq n \geq 1$; Tomkin et al., 2003).

In most studies of this type, it would be common practice to use the terrace strath as the baseline for modeling efforts, since a strath defines the depth of bedrock valley incision at that time. However, straths are rarely exposed in the Garonne valley, so terrace treads are used instead. The assumption is made that treads have the same slope as straths, and differ solely due to a relatively uniform thickness of sediment cover. Hence terrace treads are assumed to serve as a suitable proxy for changes in bed elevation through time. Long profiles of terrace tread surfaces, as defined from field efforts and published geological maps, were therefore compared to computer-generated long profiles, so as to provide information about $m$ and $n$. $q$ is taken from data published by S.M.E.P.A.G.(1989) and UNEP/DEWA (2004), and $S$ is calculated within the model.

The numerical model used here functions by first calculating the slope at each position point along the river (points are spaced one kilometer apart). Discharge is based on mean annual discharge values obtained from data published by S.M.E.P.A.G (1989) and UNEP/DEWA (2004). Slope and discharge values are then input into equation (2), as described above. Figure 10 shows the initial long profiles, slope and discharge values used in the model. The values calculated at each point are then subtracted from the initial values for the first time-step, and from the preceding values for each subsequent time-step, producing a slowly incising profile. The $K$ used for this model was 0.3.

Numerical simulations were confined to the 251 km stretch of river between Cazeres and Langon (Figure 11), so as to avoid complications from the confluence between the Garonne and Dordogne Rivers at Bordeaux, and from the Gironde Estuary and influences of
Figure 10. Initial conditions. A.) Initial long profiles of the 3rd terrace complex (yellow), 2nd terrace complex (green), 1st terrace complex (red), and floodplain complex (blue). B.) Initial slope of the 3rd terrace complex. C.) Discharge as inputted into the model is the result of tributaries downstream. After S.M.E.P.A.G., 1989.
Figure 11. Focus area map. Focus area map showing the extent of the modeled reach, the middle Garonne, highlighted in light blue. Red boxes represent the locations of the three sections discussed in this study: the lower-middle, central, and upper-middle Garonne, from north to south, respectively.
sea level. This length is considerably longer than other rivers modeled in his manner. Each kilometer along this stretch represents a single data point, such that the total population N=251.

Two criteria are used to evaluate the success of this model. First, is there reasonable error? Second, is there a pattern to the residuals? Reasonable error and randomly distributed residuals would indicate the model successfully describe long profile evolution, whereas failure to achieve either of these criteria would indicate the model does not to adequately describe the evolution of the Garonne river long profiles.

To answer the first question, the standard deviation (σ) was calculated for each set of m and n values between the 3rd terrace and floodplain complexes, and between the 3rd and 2nd terrace complexes, and the 3rd and 1st terrace complexes for select m and n values. A similar method of error calculation was used by Snyder et al. (2003) in their investigations of channel morphology in tectonic settings. The standard deviation (σ) was calculated as follows:

$$\sigma = \sqrt{\frac{\sum r^2}{N-p}}$$  \hspace{1cm} (6)

Where:

- \(\sigma\) = standard deviation of the residuals
- \(r\) = residual (simulated value – expected value)
- \(N\) = population size (251)
- \(p\) = fit parameters (2, corresponding to \(m, n\))

Calculations stopped only when the standard deviation began to increase, so as to get the best fit to the target profile. For this model, a perfect fit would produce \(\sigma=0\), using the above definition. However, error is inherent to the model as a result of error in its inputs. Assuming no error in the identification of mapped terrace elevations and the downstream tracing of long profiles, error arises chiefly from inaccurate spot elevations and discharge measurements. In this case, the impact of incorrect spot elevations would be insignificant when
compared to discharge error. There is, however, no way to calculate error in discharge measurements, as available data consists of one statistical measure of discharge per gauging station. A key assumption made regarding discharge is that current discharge is representative of past discharge.

Some caveats for the use of standard deviation as a measurement of error can be stated as follows. First, measurements of standard deviation assume results are normally distributed. This is clearly not the case when there is serial correlation, or the correlation of a variable with itself over successive time intervals. Fortunately, the presence of serial correlation also indicates failure of the model, which we also test for here. Second, use of standard deviation in this manner assumes that all data points are independent of one another, which is not entirely true, as the slope calculations used in the model link neighboring nodes. However, independence between nodes is increased when comparing distant nodes. This creates an overall independence along the river, so for our purposes, data independence is an acceptable assumption.

To determine if a generated profile produced a reasonable fit to the target terrace complex, a 68% confidence interval was chosen, corresponding to a critical value of 0.4677 and one standard deviation from the mean (McClave and Sincich, 2002). Critical values are computed as follows:

\[
C = \frac{\sigma}{\sqrt{N}}
\]  

(7)

Where:

\[
C = \text{critical value}
\]

\[
\sigma = \text{standard deviation of the residuals}
\]

\[
N = \text{population size (251)}
\]
If the model-generated profile produced a reasonable fit, 68% of the 251 points modeled would have simulated elevations within one standard deviation of the corresponding elevation points on the target terrace. Again, this method was utilized by Snyder et al (2003), though they chose a 95% confidence interval for their simulations.

The second question is more difficult to quantify as there is no single appropriate statistic to measure serial correlation. Patterns in residuals indicate serial systematic error within the model and that the model is not functioning properly. As with the standard deviation measurement, most tests for systematic error require that neighboring points are independent of each other, which is not entirely true for this case, as described earlier. However, as a means to evaluate serial systematic correlation within the programs results, a Durbin-Watson statistic was calculated for each set of $m$ and $n$ values following Draper and Smith (1998), as follows:

$$DW = \frac{N}{\sum_{i=1}^{N} (r_i)^2} \frac{N}{\sum_{i=1}^{N} (r_i - r_{i-1})^2}$$

Where:

- $DW$ = Durbin-Watson statistic
- $r$ = residual (simulated value – expected value)
- $N$ = data points in population (N)

For this river system, the Durbin-Watson statistic is too conservative, because neighboring nodes are linked by slope. However, when used in conjunction with an analysis of large-scale patterns within the residuals, the Durbin-Watson statistic provides a qualitative measurement of errors associated with serial correlation. In this case, the larger the Durbin-Watson statistic is, the less serial correlation.

True numerical ages for simulated terraces are unknown, as they are for terraces mapped in the field and from geological maps. Hence, for numerical simulations, time is
treated as a relative term, and the “time” calculated in simulations is the number of time-steps*time-step size (dt). Assuming steady rates of erosion over long periods of time, increments of “time” are considered to be proportional to the vertical distance between them.

**Geochronology**

Thirty samples were collected for purposes of optically stimulated luminescence dating (OSL) during the summer of 2004 and the summer of 2002. Sample locations were chosen based on their stratigraphic position within the terrace system and their position along the profile of the river, so as to get a representative perspective of the basin. Ten samples will be submitted for processing, but results will not be available for this thesis. Ultimately, these samples should provide numerical ages of terraces of the Garonne that can then be compared with numerical model results. In the absence of numerical ages, relative ages are known from cross-cutting relationships, and from geometric relations between terrace surfaces, the modern floodplain, and the Gironde Estuary.
RESULTS

Field Results

Three major, continuous and traceable terrace complexes, in addition to the modern floodplain complex, were mapped for the river between Cazeres and the mouth of the Gironde Estuary within the Bay of Biscay. Each terrace complex occurs over a finite elevation range that does not overlap with adjacent complexes and is bounded by major scarps on the order of 10 m or more. Such complexes are considered here to be the result of lumping minor, individual terraces of similar elevations, and the terrace complexes mapped here may represent tens of thousands of years to hundreds of thousands of years. Higher, heavily dissected terraces were not considered in this study, but could be mapped and traced in future research projects.

In the upper-middle Garonne, three major terrace complexes in addition to the floodplain were mapped, each representing numerous individual surfaces. The modern floodplain is a continuous, undissected surface with an elevation range of 225 m above mean sea level (a.m.s.l.) at Cazeres to 145 m a.m.s.l. at Portet sur Garonne, just upstream of Toulouse. The lowest, or first terrace complex, is separated from the floodplain by a 10 m scarp and ranges in elevation from 241-155 m a.m.s.l. at Cazeres and Portet sur Garonne, respectively. Again, this surface is not dissected, though is significantly wider than the modern floodplain. The second terrace complex is separated from the 1st terrace complex by a 10 m scarp and again shows little dissection. Its elevation at Cazeres is 251 m a. m.s.l. and at Portet sur Garonne is 163 m a.m.s.l. The third terrace complex ranges in elevation from 278 m a.m.s.l at Cazeres to 198 m a.m.s.l. at Portet sur Garonne. It is heavily dissected and is separated from the 2nd terrace complex by a 30 m scarp. All complexes and the floodplain maintain a slope of ~2m/km (.002) throughout this section. Figure 12 provides an example of the map distribution
Figure 12. Upper-middle Garonne geologic map and cross-section.
(A). Geologic map from BRGM plate XX-45 (1:50,000), with terraces identified in this study as shown. (B). Cross-section location as in (A), with elevations obtained from SRTM digital elevation models using the software Rivertools.
of major terrace complexes for the area of Carbonne, just north of Cazeres, and illustrates a
typical valley cross-section for this same part of the study area.

The upper-middle Garonne floodplain complex includes both overbank fines and
channel gravels and sands. Overbank sediments, where present, are typically >3 m thick,
strongly bioturbated, banded by iron, and have little clay development. Channel sands and
gravels are >10 m thick, with clasts up to 0.2 m diameter surrounded by a very coarse sand
matrix. Figure 13 is a typical measured section for floodplain sediments of the middle Garonne.
The sand and gravel deposits of the first terrace complex are >8 m thick and overlain by >2 m
of strongly cohesive, silty overbank deposits showing small (1-2 mm) iron nodules, and shells
(0.5-2 cm diameter). Gravels (>0.3 m diameter) are supported by a very coarse sand matrix and
show dune-scale trough cross-bedding. Granites within this complex are strongly weathered
and break with little force. Figure 14 is an example of a 1st terrace complex measured section
for the middle Garonne and Figure 15 shows typical outcrops from this complex and the
floodplain complex. The 2nd and 3rd terrace complexes show increasing degrees of soil
development, with peds from the 3rd terrace complex displaying blocky structures and gravels
supported by a clay rich matrix.

The central Garonne also has three major terrace complexes in addition to the
floodplain complex that could have been split, but for this purpose have been lumped to
represent distinct time periods. The modern floodplain descends 102 m between Portet sur
Garonne and Agen and is separated from the first terrace complex by a 10 m scarp. The first
terrace complex of the central section is relatively undissected, but more so than its upper-
middle counterpart. It descends 105 m over the stretch and is separated from the 2nd terrace
complex by a 10 m scarp. The 2nd terrace complex is more dissected than the first and descends
Heavily bioturbated silts and sands
Minor soil development
Small (<1 cm diameter) shells

Medium-coarse grained sands
Dune-scale trough cross-bedding

Iron-rich clays

Sands and gravels (<10 cm diameter)

--- Water Line

Figure 13. Floodplain complex measured section. Measured section from floodplain complex with unit thicknesses shown in meters. This is typical of floodplain sediments, with variations seen mainly in soil and overbank thicknesses.
Figure 14. First terrace complex measured section. Measured section of first terrace complex sediments from upper-middle Garonne with unit thicknesses shown in meters. Variations in first terrace sediments are chiefly in soil and overbank sediment thicknesses.
Figure 15. Images of middle Garonne sediments. A.) Picture of typical middle Garonne floodplain deposits. At this site, the 2+ meters of gravels (shown in inset) are overlain by 3 meters of overbank fines and sands, and underlain by 3+ more meters of gravels which extend below the lake surface.

B.) Picture of typical middle Garonne 1st terrace sediments. At this site, the dune-scale trough-cross-bedded gravels average 5 meters thickness above the talus. There is little soil development at this site.
103 m between Portet sur Garonne and Agen. A 30 m scarp marks the boundary between the 2\textsuperscript{nd} and 3\textsuperscript{rd} terrace complexes. The 3\textsuperscript{rd} terrace complex is considerably dissected and descends 108 m over this stretch of river. All three complexes and the floodplain maintain a steady slope of \(~1.4\) m/km (0.0014). Figure 16 provides an example of the map distribution of major terrace complexes for the area of Saint Nicolas de la Grave, south and east of Agen (upstream), and illustrates a typical valley cross-section for this same part of the study area.

Sand and gravel units from the central Garonne floodplain complex are typically \(<7\) m thick with large pebbles (\(<0.2\) m diameter) in a medium-coarse sand matrix showing both direction and dune-scale trough cross-bedding. Sand lenses may be strongly cemented in places, typically when in contact with overbank sediments, Overbank fines are hard and slightly blocky, with shells (0.5-2 cm diameter) and iron nodules (1-2 mm diameter) not unusual. Thicknesses of these sediments are generally \(<~3\) m. In contrast, overbank deposits from the first terrace complex show dense iron mottling, and a sandy/loam texture. Gravels (\(<0.2\) m diameter) have distinct clay skins and are supported by a medium sand matrix. Second and third terrace complexes of the central Garonne are similar to those of the 2\textsuperscript{nd} and 3\textsuperscript{rd} terraces of the upper middle section, though soils tend to be darker and show more iron mottling.

The three terrace complexes and modern floodplain of the lower-middle Garonne display the same overall patterns as the upper-middle and central sections. The floodplain descends 33 m between Agen and Castets, and is separated from the 1\textsuperscript{st} terrace complex by a 10 m scarp. The 1\textsuperscript{st} terrace complex descends from 50 m at Agen to 21 m at Castets and shows little dissection. It is again separated from the 2\textsuperscript{nd} terrace complex by 10 m. The 2\textsuperscript{nd} terrace complex descends 28 m and shows similar amounts of dissection as in the central section.
Figure 16. Central Garonne geologic map and cross-section.
(A). Geologic map from BRGM plate XIX-41 (1:50,000), with terraces identified in this study as shown. (B). Cross-section location as in (A), with elevations obtained from SRTM digital elevation models using the software Rivertools.
A 30 m scarp separates the 2nd terrace complex from the moderately dissected 3rd terrace complex that descends 30 m between Agen and Castets. The three terrace complexes and modern floodplain maintain a slope of ~0.5 m/km (0.0005) over this stretch of river. Figure 17 provides an example of the map distribution of major terrace complexes for the area of Castets, and illustrates a typical valley cross-section for this same part of the study area.

The floodplain complex of the lower-middle Garonne has sand and gravel deposits < 8 m thick with gravels (< 0.1 m diameter) supported by a coarse sand matrix containing numerous iron concretions (<2 mm diameter). Overbank sediments are coarsely sandy with little bioturbation and soil development. The first terrace complex has gravel deposits < 8 m thick covered by a thin layer of overbank fines (~1.5 m), which are argilicious in places. Sands are coarse to very coarse. The 2nd and 3rd terrace complexes are again similar to previous sections, though some locations have a much smaller maximum gravel size (<.03 m).

Lower Garonne floodplain sediments, or those of tidal influence, are typically composed of coarse sands and may be >20 m thick, considerably thicker than any seen in the upper or upper-middle reaches. This thickness may reflect Holocene aggradation due to sea-level rise.

As stated above, Hubschman (1975) interpreted the floodplain, 1st, 2nd, and 3rd terrace complexes described here to represent the Würm (MIS 2-4), Riss (MIS 6), Mindél (pre-Riss), and Günz (pre-Riss) glacial periods, respectively, based principally on stratigraphic location and morphology. Many of the details found in Hubschman’s (1975) work are not pertinent at the scale of analysis for the present study, and his geochronological and genetic interpretations might change with new understandings of glacial-interglacial cycles. However, his descriptions of sediments and soils remain valuable. In this context, he noted the floodplain deposits are composed of large amounts of postglacial sediments but that the loess deposits here possess
Figure 17. Lower-middle Garonne geologic map and cross-section.
(A). Geologic map from BRGM plate XVI-38 (1:50,000), with terraces identified in this study as shown. (B). Cross-section location as in (A), with elevations obtained from SRTM digital elevation models using the software Rivertools.
sedimentological characteristics indicative of in-place late Würm (MIS 2-4) deposition. An alternative interpretation suggested here would be that these fine silts represent flood plain facies, rather than loess, and do not imply a late Pleistocene Würm age, but are instead Holocene to modern in age. Regardless, soils are brownish and include BC to C horizons with massive pedogenic structure. Hubschman’s (1975) “Riss” (MIS 6) terrace has many of the same features as the floodplain, including similar degrees of mineral alteration, but with a calcareous horizon at depth. This terrace also has more concretions (ferromagnesian) and more soil development. The Mindél (pre-Riss) terrace has thicker soils, with abundant concretions and bioturbation. Gravels in the soil profile lie within a soil/sand matrix and show extensive mineral alterations. The high Günz (pre-Riss) terrace shows the same sequences as the lower Mindél terrace, but with a greater degree of soil development and mineral alteration. The soils are more calcareous, show more warping, and are more strongly colored. Sand grains lower in the sequence show in-place alteration and gravels have higher degrees of mineral alteration tending towards a calcareous crystal structure (Hubschman, 1975).

**Construction of Long Profiles**

Long profiles created for the Garonne River floodplain complex and adjacent three terrace complexes are shown in Figure 18, and reveal uniform separation of ~10 between the floodplain, 1st and 2nd terrace complexes and ~30 m between the 2nd and 3rd terrace complexes.

**Numerical Modeling Results**

An overview of standard deviations for all \( m \) and \( n \) values evaluated in this study \((1 \geq m \geq -0.5 \text{ and } 2 \geq n \geq 0)\) reveals a general trend of decreasing standard deviation as values approach \( m=n=0 \) (Figure 19). Critical values, \( C \), within one standard deviation (\( \sigma \)) of the mean \((C = 0.4677)\) are expected for these simulations and should more than account for error in
elevation and discharge measurements. It should be noted that calculations of the variation in downstream discharge between mean annual values and those for 5-year flooding events, or bank-full values, produces a 2:3 relationship instead of a 1:1 relationship (Table 1). This means that there is too much variation in the discharge input used in this model, and therefore the model underpredicts the sensitivity of the system to discharge. Consequently, the actual \( m \) value is between 0.25 and 0.5 greater than the results indicate. However, to be consistent, uncorrected values are used in figures and text here.

Table 1. Discharge Corrections

<table>
<thead>
<tr>
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<th>Mean Annual Discharge (( Q_{ma} ))</th>
<th>5-yr. Flood Discharge (( Q_5 ))</th>
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<tbody>
<tr>
<td>Toulouse (Upstream)</td>
<td>200 m(^3)s(^{-1})</td>
<td>2450 m(^3)s(^{-1})</td>
</tr>
<tr>
<td>Langon (Downstream)</td>
<td>600 m(^3)s(^{-1})</td>
<td>4700 m(^3)s(^{-1})</td>
</tr>
<tr>
<td>Upstream/Downstream</td>
<td>( Q_{ma} / Q_5 )</td>
<td>0.33</td>
</tr>
<tr>
<td></td>
<td>0.63</td>
<td>0.52</td>
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</tbody>
</table>

The first hypothesis, or unit stream power/shear stress case (\( 2m = n \) for \( 2.5 \geq n \geq 0.66 \)) produced consistently poor fits for all \( m \) and \( n \) tested. Four \((m, n)\) pairs are used here to illustrate the fit of this model to the Garonne River, (0.25, 0.75), (1, 2), (0.75, 1.5) (Figure 20), and (0.5, 1) (Figure 21). The most common \((m, n)\) combination, (0.5, 1) had standard deviations (\( \sigma \)) ranging from 13.39 for the 3\(^{rd}\) to 2\(^{nd}\) terrace complexes to 20.47 for the 3\(^{rd}\) terrace to floodplain complexes and values of \( C \) ranging from \( C = 0.845 \) to 1.292, respectively. Such values are well outside the acceptable one standard deviation error (\( C = 0.4677 \)) and indicate a generally poor fit for all simulated terrace complexes. This case also produced strongly patterned residuals, with average Durbin-Watson statistic of 0.0006, and qualitative analysis.
Figure 18. Long profiles. Long profiles of the floodplain complex (red) and 3rd (black), 2nd (green), and 1st (blue) terrace complexes.
Figure 19. Error map for all $m$ and $n$. Error Map (A) and 3D plot (B).

Contours are colored to reflect the amount of deviation between generated and target profiles for all $n$ and $m$ tested.

a. Transport-limited ($n=1, m=0$)
b. Shear stress ($n=1, m=.5$)
c. Best fit ($n=0, m=0$)
d. Total stream power ($n=.25, m=.25$)
Figure 20. Results for unit stream power/shear stress simulations (various $m,n$).
Results of numerical simulations for unit stream power/shear stress model
when A.) is $(0.25, 0.75)$, B.) is $(0.75, 1.5)$, and C.) is $(1, 2)$. Long profiles
are on the left and residuals are on the right. The 3rd terrace is shown in
yellow, the modern floodplain in red, and the generated terrace in green.
Residuals are shown in blue and correspond to the offset
between the red and green profiles.
Figure 21. Results for unit stream power/shear stress simulations ($m = 0.5$, $n = 1$). Results are shown with long profiles on left and residuals on right. The 3rd terrace is shown in yellow, the target terrace in red, and the generated terrace in green. Residuals are shown in blue and correspond to the offset between the red and green profiles.
indicates rates of erosion are overestimated in the upstream reaches and underestimated in downstream reaches, producing a downstream convergence of the simulated terrace with the original terrace. A comparison of $m=0.5$ and $n=1$ with other $m$ and $n$ pairs used here reveals a general trend of increasing downstream convergence with increasing $n$ and $m$ values. Correspondingly, as this convergence increases, so does the error, with the most convergent case, $(1, 2)$ having the greatest error, $C = 18635$ for the simulation between the 3$^{rd}$ terrace complex and floodplain complex. Table 2 gives values of $\sigma$, $C$, and Durbin-Watson statistics, for all unit stream power/shear stress simulations.

The transport-limited case also produced consistently poor fits for all combinations of $m$ and $n$ within the model range $(0 \geq m \geq -0.5$ and $2 \geq n \geq 1)$. For purposes of illustration, four $(m, n)$ combinations are compared, $(0, 2), (-0.5, 1), (-0.5, 2)$ (Figure 22), and $(0, 1)$. The best fit transport-limited position of $(0, 1)$ is plotted in Figure 23. Standard deviations ($\sigma$) for this case were $18.0, 22.81$, and $27.93$ for simulations between the 2$^{nd}$ and 3$^{rd}$ complexes, 1$^{st}$ and 3$^{rd}$ complexes, and floodplain and 3$^{rd}$ complexes, respectively. These standard deviations correspond to critical values of $C = 1.136, 1.440$, and $1.763$, respectively, all of which are significantly greater than the expected value of $C = 0.4677$. For all $m$ and $n$ within this model range, residuals show that upstream sections of simulated profiles overestimate rates of erosion, whereas rates are underestimated in the downstream reaches. Durbin-Watson statistical values are well outside the limitations of the test and indicate strong serial correlation. Like the unit stream power/shear stress case, this model produces convergence between the terraces in a downstream direction, with $(-0.5, 2)$ converging the most and having the worst fit. Standard deviations, critical values, and Durbin-Watson values are listed for the above $(m, n)$ combinations and others within the range of this test in Table 2.
Table 2. Numerical Modeling Results

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<th></th>
<th>Unit stream power/shear stress</th>
<th>Transport limited</th>
<th>Total Stream Power</th>
<th>m &gt; n</th>
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<tr>
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<td>Durbin-Watson Statistic</td>
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Figure 22. Results for transport limited simulations (various $m$, $n$). Results of numerical simulations for the transport limited hypothesis. A.) is (0, 2). B.) is (-0.5, 1). C.) is (-0.5, 2). Long profiles are on the left and residuals on the right. The 3rd terrace is shown in yellow, the modern floodplain in red, and the generated terrace in green. Residuals are shown in blue and correspond to the offset between the red and green profiles.
Figure 23. Results for transport-limited simulations ($m=0, n=1$). Long profiles are on left and residuals on right. The 3rd terrace is shown in yellow, the target terrace in red, and the generated terrace in green. Residuals are shown in blue and correspond to the offset between the red and green profiles.
The total stream power hypothesis, where \( m=n \) was more successful than any hypotheses tested. \( m=n \) combinations such as \( m=n=1 \), \( m=n=0.75 \), \( m=n=0.5 \) (Figure 24), and \( m=n=0.25 \) (Figure 25) produce decent fits, with \( m=n=0.25 \) producing the second best fit of all \( m \) and \( n \) combinations tested. C values for \( m=n=0.25 \) ranged from 0.149 to 0.188, sufficiently low to produce a generated profile that closely resembles the target profile. Residuals show a near random distribution, with the Durbin-Watson statistic averaging 0.00837, and appear to be non-serially correlated. Table 1 illustrates these values.

In addition to the above three models, a situation in which \( m>n \) was evaluated as well. In this simulation, \( m>n \) is illustrated by \((m, n)\) of \((0.75, 0.5)\), \((1, 0.75)\), and \((1, 0)\) (Figure 26). In all cases, divergence of terrace surfaces occurs in the downstream direction, with underestimation of incision upstream and overestimation of incision downstream, though it is not consistently significant; this scenario produced three results within one standard deviation (\( \sigma \)) of the mean, as many as the total stream power hypothesis. The best fits from this case are those where \( m \) is only \( 0.25 > n \), and error increased as the distance between \( m \) and \( n \) grew. The most divergent case is that of \((1, 0)\), and the least divergent is \((0.75, 0.5)\). Table 1 gives critical values, standard deviations, and the Durbin-Watson statistic.

When \( m=n=0 \), there is a failure of the model or slope and discharge independent incision. For this situation, \( C=0.1137, 0.1160, \) and 0.1178, for calculations between the 2\textsuperscript{nd} and 3\textsuperscript{rd} terrace complexes, 1\textsuperscript{st} and 3\textsuperscript{rd} complexes, and floodplain and 3\textsuperscript{rd} complexes, respectively, with an average value of \( C = 0.116 \). This value is significantly less than \( C = 0.4677 \), the expected error for one standard deviation (\( \sigma \)). Accordingly, profiles of simulated terraces closely match the target terrace or floodplain by showing no convergence or divergence in the downstream direction. In addition, residuals show no discernable pattern (Figure 27) and have
an average Durbin-Watson statistic of 0.0194, higher than any other tested $m$ and $n$. Thus, this combination of $m$ and $n$ produced the best fit between model simulations and field data.
Figure 24. Results for total stream power simulations (various \( m, n \)).

A.) is (1, 1), B.) is (0.75, 0.75), C.) is (0.5, 0.5).

Long profiles are on the left and residuals on the right. The 3rd terrace is shown in yellow, the modern floodplain in red, and the generated terrace in green. Residuals are shown in blue and correspond to the offset between the red and green profiles.
Figure 25. Results for total stream power simulations \((m=n=0.25\)). Long profiles are on left and residuals on right. The 3rd terrace is shown in yellow, the target terrace in red, and the generated terrace in green. Residuals are shown in blue and correspond to the offset between the red and green profiles.
Figure 26. Results for simulations when $m>n$ (various $m,n$).
A.) is (0.75, 0.5). B.) is (1, 0.75). C.) is (1, 0). Long profiles are on the left and residuals are on the right. The 3rd terrace is shown in yellow, the modern floodplain in red, and the generated terrace in green. Residuals are shown in blue and correspond to the offset between the red and green profiles.
Figure 27. Results for simulations when $n = m = 0$. Long profiles are on left and residuals on right. The 3rd terrace is shown in yellow, the target terrace in red, and the generated terrace in green. Residuals are shown in blue and correspond to the offset between the red and green profiles.
DISCUSSION

Empirical Data

The floodplain and three major mappable terrace complexes of the middle Garonne were traced for 251 km in a downvalley direction from Cazeres to Castets. Each terrace complex is interpreted to represent a significant period of time over which lateral migration, cutting of bedrock straths and deposition of fluvial sand and gravel predominated over bedrock incision and valley deepening. Minor terraces can be identified within the younger large-scale complexes, terraces 1 and 2, but were not addressed in this study; the major complexes considered here are separated from adjacent complexes by major scarps with relief in excess of 10 meters. Assuming uniform thicknesses of sediment cover underneath each terrace surface, the relief between successive terrace surfaces is interpreted to reflect the depth of bedrock incision that occurred between each major episode of lateral migration, strath formation and fluvial deposition.

As previously discussed, the uniform nature of the terrace complexes of this river rendered the model unable to produce meaningful numerical age estimates for its three major terrace complexes. Future research should focus on the development of an independent geochronological framework from OSL or cosmogenic radionuclide dating. For the purposes of this study, general age estimates for terrace complexes in any valley cross-section can be made if long-term average rates of incision are known, and if one then assumes a constant rate of bedrock valley incision. Neither rates of uplift or of incision can be determined from the study area proper, however, studies from elsewhere in lowland Europe can be useful in this context. Previous work on the Thames, Rhine-Meuse (Maas), Loire-Allier and others (e.g. van den Berg, 1996; Westaway et al., 2002; Bridgland, 2000; Brocard et al., 2003; Westaway et al.,
2003; Westaway, 2004; Wallinga et al., 2004) use river terraces to calculate incision rate and infer uplift. They indicate a relatively stable average regional uplift/incision rate of 0.1 to 0.2 mm yr\(^{-1}\) through the Pleistocene, but these values vary significantly both geographically and temporally. For example, incision rates of the Rhine-Meuse system are interpreted to have increased since the late Pliocene from 0.03 mm yr\(^{-1}\) to 0.61 mm yr\(^{-1}\) (Bridgland, 2000).

These European rates are considerably higher than what has been suggested for large parts of the North American continental interior, including the Rocky Mountains and Great Plains, where the highest rates of incision are ~0.15 mm yr\(^{-1}\) (Dethier, 2001), but significantly less than values obtained for the Clearwater River in the tectonically-active Olympic Mountains of Washington State, where incision can be as great as ~0.9 mm yr\(^{-1}\) (Pazzaglia and Brandon, 2001; Tomkin et al., 2003), so they may be useful estimates. However, unlike the streams of the Olympic and Rocky Mountains, the Garonne is located within an inactive orogen.

Taking incision rates of 0.1-0.2 mm yr\(^{-1}\), as derived from studies of European rivers, and assuming steady rates through time, preliminary age estimates can be assigned to each of the major terrace complexes in this study. Assuming the lower value of 0.1 mm yr\(^{-1}\), terrace 3 of this study is estimated to have formed ca. 500 kyr BP (MIS 13), terrace 2 would have formed ca. 200 kyr BP (MIS 7), and terrace 1 would have formed 100 kyr BP (MIS 5). Assuming higher values of 0.2 mm yr\(^{-1}\), terrace 3 of this study is estimated to have formed ca. 250 kyr BP (MIS 8), terrace 2 would have formed ca. 100 kyr BP (MIS 5), and terrace 1 would have formed ca. 50 kyr BP (MIS 3) (Imbrie et al., 1984).

Although common practice among European workers (e.g. Hubschman, 1975; Maddy, 1997; Bridgland, 2000; Maddy et al., 2001; Westaway et al., 2002; Maddy et al., 2003; Westaway et al., 2003; Westaway, 2004), it is not possible with this data to infer whether
periods of strath cutting and deposition, or conversely, periods of bedrock valley incision, correspond with glacial or interglacial climatic conditions. However for comparative purposes, age estimates provided above would place periods of deposition and periods of terrace formation within both glacial and interglacial periods.

**Numerical Modeling Data**

Numerical modeling results provide insights into long profile evolution. As shown in Figure 17, plots of long-profiles of the floodplain and terrace complexes confirms that they remain evenly spaced for the entirety of the middle Garonne study reach. With this in mind, it is not surprising that the unit stream power (shear-stress) case does not work, regardless of the alluvial vs. bedrock nature of the channel. This model would predict that terraces should converge in a downstream direction, because it places more importance on slope than discharge, and slope decreases in the downstream direction. The same argument can be made for the transport-limited case, which also predicts convergence of terrace profiles downstream. However, the stream did incise in both cases, indicating that the threshold of basal shear stress was overcome and that the stream had sufficient power to transport available sediment.

More interesting are the total stream power case, which places equal dependence on slope and discharge, and the situation when \( m > n \), which predicts divergence in a downstream direction by emphasizing the importance on discharge over slope. All total stream power scenarios tested here (see Table 2), produced consistently better results/fits for this river than unit stream power/shear-stress, transport-limited, or \( m > n \) models. There is no presently available model that describes a relationship in which \( m > n \), but this scenario produced better fits than the more generally accepted shear-stress/unit stream power and transport-limited models, so such a scenario warrants further study.
The best-fit case consists of \( m = n = 0 \), and is the result of uniform rates of incision throughout the 250 km study reach, rates that are independent of both slope and discharge. This situation is the null hypothesis of this study and as such, it means little except (a) that none of the other models tested inform us about the hydraulics behind incision in this river, and (b) that incision is independent of slope and discharge. Fortunately, total stream power and \( m > n \) models were somewhat successful and can be used to interpret incision in this system.

One important thing that can be taken from these modeling results is that we now know what parameters are not responsible for long profile evolution in the Garonne system. With that said, some other mechanism must be the driving force for long-term trends in valley incision.

**Linking Empirical Data and Model Data**

Given that long profiles remain evenly spaced through the length of the study reach, over a distance of 251 km, it can be inferred that long-term average rates of bedrock incision are relatively uniform over this distance as well. Additionally, from long profile data, it is safe to say that, while this is the Aquitaine sedimentary basin, it has not been subsiding during the time period over which terraces formed. The majority of subsidence in this region is thought to have occurred during the Mesozoic as a response to plate movements (Brunet, 1984). From data presented here, it seems clear that the Garonne drainage now behaves as an upland erosional landscape.

From the numerical modeling exercises, it is clear that long profile evolution is not explained by either a stream-power (shear-stress) or transport-limited model. One possible explanation may be non-uniform incision, which has not been explored here, but could be a topic of future research. The majority of uplift in France is now centered around the Massif Central, where there are greater crustal thicknesses and higher heat flow. However, uplift rates
for the Massif Central are not steady and can be correlated with volcanism, which was limited during the Middle Pliocene and early Late Pleistocene (Westaway, 2002; Westaway, 2004). Additional uplift during this time is attributed to isostatic adjustment to the Pyrenean orogeny along the North Pyrenean fault zone, which created both steepened Pyrenean slopes and dropped the central regions of the fold belt in excess of 1000 m (Bourrouilh et al., 1995). This unsteady uplift, both spatially and temporally may explain why the unit stream power/shear stress, transport limited, total stream power, and \( m > n \) simulations did not work as well as the null hypothesis.

The idea of spatially variable and temporally unsteady uplift may have significant effects within this river system and the model outcomes. It is not unreasonable to assume that tectonic uplift along the Pyrenean front has been more significant than in the distal parts of the basin. With that said, if uplift is included into the model and varied spatially, the disagreement between the simulated profiles and real profiles may be significantly diminished. Both the unit stream power/shear-stress and transport-limited models overestimated incision in the upstream reaches and underestimated incision in the downstream reaches. By simulating increase uplift rates in the upstream reaches, relative to farther downstream, this effect could be diminished.

Another shortcoming that might be addressed in future model simulations would be unsteadiness in rates of uplift. As discussed, there are two potential sources of uplift with the region, the Pyrenees and Massif Central. In both cases, the majority of tectonic activity is confined to a specific time, determined via fission track (Shoemaker, 2000) and (Westaway, 2004). Since those major events, they have been relatively inactive, indicating that whatever uplift has happened has slowed appreciably. One potential correction that could be made to this model is an incorporation of uplift that can be varied temporally and spatially.
An important consideration in considering the evolution of any fluvial system is how well it conforms to the concept of a steady state system. A steady-state system implies that the river is incising at the same rate as uplift is taking place (Bull, 1991). This idea has been questioned recently by Whipple (2001), who suggested that a steady-state system (or graded-stream) is not realistic during times of climatic instability due to the effects climate variations have on river hydrology. The climate of France is, as discussed, very well understood, and detailed studies have revealed variability within the past 500 kyrs in the form of six major interglacial periods (Reille and de Beaulieu, 1995; Roger et al., 1999; Raynaud et al., 2005). It may be appropriate, then to include parameters that can account for unsteadiness in discharge regimes (glacial vs. interglacial) within a future modeling effort, such as employed by Baldwin et al. (2003) in their examination of the effects of postorogenic decay on fluvial systems.

Finally, an alternative to the theory that tectonic uplift is the driving force behind incision in this system, is that incision in the Garonne may be explained by differences in lithology, rock strength, spatially-variable stresses, grain size, etc. It would be difficult to test some of these hypotheses, but recent work by Lisle et al. (2000), and Sklar and Dietrich (2001), has shown some promising results with respect to sediment and rock strength that could be utilized in further studies of this system.
CONCLUSIONS

The Garonne River heads in the moderately glaciated Pyrenees of northernmost Spain and southwestern France, traverses the Aquitaine Basin, then discharges into Atlantic Ocean via the Gironde Estuary/Bay of Biscay. The Garonne presents a unique opportunity to study the mechanisms and controls on long-term incision and terrace formation within a large-scale fluvial system. From field data, three terrace complexes were identified and traced over a downvalley distance of >251 km, from the base of the Pyrenees to the onset of tidal influences. Each terrace complex occurs over a finite elevation range that does not overlap with adjacent complexes and is bounded by major scarps on the order of 10 m or more. Each terrace complex is interpreted to represent a significant period of time over which lateral migration, cutting of bedrock straths and deposition of fluvial sand and gravel predominated over bedrock incision and valley deepening. Assuming uniform thicknesses of sediment cover underneath each terrace surface, the relief between successive terrace surfaces is interpreted to reflect the depth of bedrock incision that occurred between each major episode of lateral migration, strath formation and fluvial deposition.

Following previous work, a numerical model was created to test the dependence exerted by slope and discharge on incision within the Garonne River system. Hypotheses tested include detachment-limited, transport-limited, and total stream power relationships, as well as a null hypothesis that specifies incision to be independent of discharge and slope. Each model has a specific range of exponents for discharge and slope (m and n values, respectively) within the overall incision equation. Error and fit for each model formulation were evaluated using statistical tests. Results from the detachment-limited and transport-limited models were all unacceptable, whereas the total stream power model produced three acceptable combinations of
\(m\) and \(n\). However, the best-fit result for this system is the null hypothesis. This result means little except that none of the models tested here are able to describe the hydraulics behind incision in this river, or that valley incision is independent of slope and discharge.

One possible mechanism to explain incision in this system is unrecognized uplift in southwestern France. General age estimates for terrace complexes in any valley cross-section can be made if long-term average rates of incision are known, and if one then assumes a constant rate of bedrock valley incision. Previous work on the Thames, Rhine-Meuse (Maas), Loire-Allier and others indicate a relatively stable average regional uplift/incision rate of 0.1 to 0.2 mm yr\(^{-1}\) since the Pleistocene. Using incision rates estimated for these systems, the 3\(^{rd}\) terrace complex formed 500-250 kyr BP, the 2\(^{nd}\) terrace complex formed 200-100 kyr BP, and the 1\(^{st}\) terrace complex formed 100-50 kyr BP. From this data it is impossible to link strath cutting or deposition to specific glacial or interglacial cycles, or specific climatic conditions.

Additionally, it is reasonable to assume that tectonic uplift along the Pyrenean front has been more significant than in the distal parts of the basin. Therefore, if uplift is included into the model and varied spatially, the disagreement between the simulated profiles and real profiles may be diminished for the more commonly accepted transport-limited, detachment-limited, and total stream power hypotheses. An alternative theory to that of incision-driven uplift is that differences in rock strength, spatially-variable stresses, grain size differences, etc. play a major role in bedrock incision. These may be difficult to quantify in a system the size of the Garonne, but such variations could be a focus of further study. Finally, numerical age constraints on periods of deposition and incision can provide the means for independent estimates of incision and uplift rates in this region and provide for correlation between the timing of terrace formation and climate and sea-level changes.
REFERENCES


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VITA

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