Of ice and water: Quaternary fluvial response to glacial forcing

Stéphane Cordier[a], Kathryn Adamson[b], Magali Delmas[c], MarcCalvet[c], Dominique Harmand[d]

[a]Département de Géographie-UMR 8591 CNRS-Université Paris 1-Université Paris Est Créteil, 61 avenue du General de Gaulle, 94010 Créteil cedex, France. stephane.cordier@u-pec.fr

[b]School of Science and The Environment, Manchester Metropolitan University, M1 5GD Manchester (UK). k.adamson@mmu.ac.uk

[c]Université de Perpignan-Via Domitia, UMR 7194 HNHP, 66860 Perpignan Cedex, France. magali.delmas@univ-perp.fr, calvet@univ-perp.fr

[d]Laboratoire LOTERR, Université de Lorraine, site Libération, BP 13387, 54015 Nancy, France. dominique.harmand@univ-lorraine.fr

Abstract

Much research, especially within the framework of the Fluvial Archives Group, has focused on river response to climate change in mid-latitude non-glaciated areas, but research into the relationships between Quaternary glacial and fluvial dynamics remains sparse. Understanding glacial-fluvial interactions is important because glaciers are able to influence river behaviour significantly, especially during glacial and deglacial periods: 1) when they are located in downstream from a pre-existing fluvial system and disrupt its activity, leading to hydrographical, hydrosedimentary and isostatic adjustments, and 2) when they are located upstream, which is a common scenario in mid-latitude mountains that were glaciated during Pleistocene cold periods. In these instances, glaciers are a major water and sediment sources. Their role is particularly significant during deglaciation, when meltwater transfer towards the fluvial system is greatly increased while downstream sediment evacuation is influenced by changes to glacial-fluvial connectivity and basin-wide sediment storage. This means that discharge and sediment flux do not always respond simultaneously, and this can lead to complex fluvial behaviour involving: proglacial erosion and sedimentation, and longer-term paraglacial reworking. These processes may also vary spatially and temporally according to the position relative to the ice margin (ice proximal versus ice distal location). With a focus on the catchments of Europe, this paper aims to review our understanding of glacial impacts on river system behaviour. We examine the methods used to unravel fluvial response to ‘glacial

forcing', and propose a synthesis of the behaviour of glacially-fed rivers, opening perspectives for further research.

1. Introduction

River systems are highly sensitive to environmental changes including: tectonic, climatic, glacial, and anthropogenic forcing. Fluvial morphosedimentary records, and the natural (e.g. palaeontological) and human (archaeological) archives preserved within them, can provide valuable palaeoenvironmental information. They allow us to reconstruct environmental evolution at local to regional scales, and over modern to Pleistocene timescales. Reconstructing Quaternary river dynamics is fundamental to our understanding of present day fluvial systems because long-term Quaternary incision has shaped modern valley landscapes (Bridgland and Westaway, 2007). At the same time, the study of present day river systems makes it possible to better understand the significance of the older, Pleistocene, fluvial archives, and the relationship between catchment evolution and fluvial dynamics.

The impacts of Quaternary glacial-interglacial cycles on mid-latitude river systems have long been emphasised (e.g. Vandenberghe, 1995, 2003, 2008, 2014; Bridgland and Westaway, 2007). These climatic influences have been direct and indirect: temperature and rainfall directly control river discharge and, in many cases, erosion and sediment production. Climatic controls on the presence/absence of permafrost and the type/quantity of vegetation cover, exert an indirect control on river system behaviour. Both of these parameters influence catchment-scale water and sediment transfer, from the hillslopes to the valley floor and channel(s) (Vandenberghe, 1995, 2001). The complexity of fluvial response to Pleistocene climate change has been investigated for many decades (e.g. Sörgel, 1939; Büdel, 1977; Vandenberghe, 1995, 2003; Bridgland, 2010). Research has mainly focused on fluvial systems from Northwest Europe, which were characterized by periglacial conditions during Pleistocene cold periods. The Thames, Meuse, Somme, Rhine, and Vistula catchments have been investigated in detail, and have become established as reference areas for the reconstruction of Quaternary climate forcing on fluvial systems (e.g. Bridgland, 1994; Starkel, 1994; van den Berg, 1996; Antoine et al., 2000, 2007; Busschers et al., 2007; van Balen et al., 2010). They have provided a better understanding of hillslope-river coupling at the 100 ka timescale. However, many studies focusing on climate forcing either were performed on non-glaciated catchments/sections of valleys, or have paid little attention to the presence of glaciers in the upstream part of the catchment, as is the case for the Rhine (e.g. Boenigk and Frechen, 2006, van Balen et al., 2010). In fact, despite the evidence that glaciers covered up to 30% of the global land surface during some Pleistocene cold periods, and the fact that fluvial terraces have been identified downstream of glaciated areas for more than a century (Carney, 1907; Penck and Brückner, 1909), the relationships between glacial and fluvial dynamics have not been examined in detail except for some areas such as the United-Kingdom (Bridgland and Westaway, 2014). However, these relationships are important because: 1) the course of a river can be transformed by the damming of valleys by ice or moraines; 2) glaciers play a major role in shaping landscapes through erosion; 3) this erosion produces vast amounts of sediments that are transported downstream by rivers; and 4) glaciers are major water reservoirs that can strongly influence catchment hydrological regime.

Following from this, the glacial control on river behaviour cannot be considered as unequivocal. To effectively understand the impacts of glacial activity, it is actually important to establish the spatial relationship between glacial and fluvial systems. Two main scenarios should be distinguished:

-the first one corresponds to glacially disrupted rivers, when glaciers, and especially ice sheets, occupy a part (which can be located either in the headwaters or further downstream) of a pre-existing fluvial system ('downstream control'). This often leads to the destruction of the previously formed fluvial archives (Bridgland and Westaway, 2014). Rivers can even be obliterated completely by glaciations, as was the case, for instance, for the Scandinavian fluvial systems, for the proto-Soar/Bytham river in Great Britain (White et al., 2010, 2016; Gibbard et al., 2013) or for the Ohio system (Jacobson et al., 1988; Granger et al., 2001). This first scenario is typically found in lowlands area of northern Europe (from the UK to Germany, Poland, Ukrainia and Russia) and Northern America, which has been largely covered by ice sheets during Pleistocene cold periods. It may also be observed locally in montane areas when a glacier dams a valley. In that case, the response of the fluvial system is, however, different from lowland area, first since the glacial damming of a fluvial valley is typically a transitional phenomenon occurring at the beginning or the end of a glacial period (see below 5.1), and secondly since it affects confined systems.

-the second scenario corresponds to glaciers developing in the upstream parts of the fluvial systems ('upstream' control). Such glacially fed rivers are typical from montane areas, but can also be found in lowland areas in case of southwards drainage systems fed by meltwater from ice sheets, such as the Dnieper or the Don in Eastern Europe.

This paper examines the influence of Quaternary glacial activity both on 'disrupted' and 'glacially-fed' river systems. The first section focuses on the methods that are typically used to recognize glacial forcing in the fluvial record, in particular at the Pleistocene timescale. The key role of geochronology (Rixhon et al., this issue) and modelling in conjunction with the indispensable field-based approach (morphological and sedimentological investigations) is underlined. The second section corresponds to an extensive review of the way how the fluvial activity may be disrupted by the glaciers, especially when these are located downstream. The following sections focus more specifically on the 'upstream' control. The latter actually involves a complex pattern of fluvial response, since glaciers located in the headwaters are able to influence both the water and sediments flows. We focus in particular on glacial-fluvial interactions in glacially-fed rivers during periods of ice retreat, because this transitional period is characterised by major shifts in meltwater and sediment dynamics that control the response of the fluvial systems downstream. We then develop a review of recent research applying these methods to examine Pleistocene glacial-fluvial interactions in catchments across Europe. This allows us to assess the nature of glacial forcing on fluvial behaviour, and unravel the importance of connectivity in glacial-fluvial systems. The subsequent discussion examines whether fluvial response to glacial dynamics (and in particular glacier retreat) in different basins, is characterized by uniformity or diversity, and to open perspectives for further research.

2. Methods to study fluvial response to Pleistocene glacial changes: a multi proxy approach

Unravelling the influence of glacial activity on fluvial system behaviour, requires a good understanding of extent of the glaciers in the case of disrupted fluvial systems. For glacially-fed rivers, key parametres are characteristics and timing of water flow and sediment flux - including the possibility of short-or long-term storage in morphological depocentres (Koppes and Montgomery, 2009). Many studies have examined *either* glacial *or* fluvial system dynamics, but few have developed a coupled glacial-fluvial approach. As a consequence, there is an empirical 'grey area' in our understanding of the links between ice proximal meltwater outwash dynamics, and the typical fluvial archives recognized kilometres or tens of kilometres downstream. Bridging this gap is a key research objective.

Several methods may be used to examine glacial-fluvial interactions, and these can be broadly categorised as: morphology/sedimentology, geochronology, and modelling.

2.1 Morphology and sedimentology

High-resolution geomorphological mapping of landform asemblages is key for distinguishing between glacial, transitional, and fluvial settings, and for exploring spatiotemporal relationships between glacial and fluvial processes both for 'downstream' and 'upstream' controls (Flageollet, 2002; Bridgland and Westaway, 2014; Stange et al., 2014; Delmas et al., 2015). This distinction can be challenging, especially for the older part of the Quaternary record; where glaciers no longer exist and spatial relationships between glacial and fluvial systems are unclear; and where landforms inherited from earlier Pleistocene cold periods have been reshaped or fragmented by subsequent fluvial or slope processes. We therefore rely on a combination of morphological (e.g. identification of moraines) and sedimentological evidences as sediment structure, bedding and grain characteristics vary profoundly between glacial and fluvial settings. Detailed analysis of catchment topography allows us to examine pathways of meltwater and sediment flux and locate depocentres that may have disrupted downstream sediment transfer, and altered glacial-fluvial connectivity. The value of field mapping cannot be overestimated, as demonstrated by the Fluvial Archives Group. This approach can be enriched by the use of thematic maps, air photos, satellite remote sensing, and digital elevation models (DEMs) which have enabled landform recognition over large areas and/or where fieldwork is problematic (Wiederkehr et al., 2010; Pazzaglia, 2013).

A combination of mapping techniques, such as those outlined above, is likely to provide the most robust reconstruction of glacial-fluvial interactions. This is because, in glaciofluvial settings, the landscape is shaped by processes operating over two dimensions: 1) changes in glacier behaviour largely, though not exclusively, occur on a longitudinal profile (advance/retreat of the ice margin, downstream transfer of meltwater, ice, and sediment). 2) In contrast, the fluvial system cannot be understood without also including a significant vertical component, which is expressed by fluvial incision. This is clearly apparent in uplifted areas, including middle and high mountains that were glaciated during the Quaternary. The interaction of longitudinal and vertical processes can produce complex glaciofluvial landform assemblages. Even if morphostratigraphic correlations between glacial and fluvial landforms can be firmly established for the last glaciation, the task is more challenging for previous cold stages because landforms have been exposed to multiple phases of erosion/reworking, and are often poorly preserved.

Sediment analysis allows us to examine glacial-fluvial interactions in further detail. As in non-glaciated basins, alluvial records in glacially-fed rivers are indicative of environmental conditions at the time of deposition. This is based on sediment characteristics such as facies arrangements and structures such as periglacial deformations. Grain size may also be indicative of ice proximity, and associated changes to channel flow conditions and sediment inputs. Sediment lithology can be used to 'fingerprint' glacier and meltwater source area. This approach is especially effective where the limit between glaciated and non-glaciated areas coincides with a lithological boundary, as is the case in the Moselle catchment (Cordier et al., 2004, 2006). Other evidence can be derived from biostratigraphic markers such as malacofauna. Collectively, these analyses allow us to reconstruct sediment transportation processes and depositional context, and draw relationships with glacial (and therefore climatic) regime. However, to make meaningful correlations between phases of fluvial aggradation and erosion, glacier mass balance, and Quaternary climate changes, a robust numerical geochronology is required.

2.2 Geochronology

Several dating methods are commonly used to establish the age of Quaternary glacial and fluvial activity (Rixhon et al., this issue). Relative ages may be adequately derived from amino-acid racemisation, soil development, and biostratigraphy. The latter was proven helpful especially in the UK, where it allowed, in combination with the morphostratigraphical correlations between glacial and fluvial archives, the recognition of interglacial deposits within the fluvial sequences and so the indirect dating of glaciations (White et al., 2016, this issue). Amongst numerical dating methods, radiocarbon dating is widely used to establish the timing of fluvial changes, but it relies on the presence of organic material, may lacks in glaciated areas. Instead, cosmogenic nuclides, optically stimulated luminescence, and uranium-series techniques have become more commonly applied both to glacial and fluvial sedimentary sequences over the last few decades, so they will be briefly described here.

2.2.1. Terrestrial cosmogenic nuclide dating

In-situ terrestrial cosmogenic nuclide (TCN) dating of glacial and fluvial deposits requires various sampling and modelling methodologies, depending on the morphological, sedimentological and palaeoenvironmental context of the study. In glacial environments, TCN dating aims to reconstruct spatial fluctuations of glacier margins through timeusing erratic boulders preserved onmoraine ridges, and/orpolished surfaces located on bedrock-steps (Ivy-Ochs and Briner, 2014). To ensure reliable ages, the selected surface must be chosen with care. Boulders should only be used if their pre-glacial TCN dose has been removed ('zeroed') by erosion during glacial transport, such that the measured TCN signal is synchronous with boulder deposition in the ablation till. Sampled boulders must therefore display morphological and/or lithological evidence for long-term glacial transport.For polished bedrock surfaces, it is assumed that the layer removed by glacial erosion was thick enough (i.e. >2-3 m) to reset the TCN signal associated with the previous interglacial/interstadial. Samples should be obtained from surfaces within the main axis of paleo-ice flow, where subglacial erosion is concentrated. If this condition is not satisfied, the apparent ages can be significantly older than the genuine deglacial age, and cannot provide a reliable geochronology. Conversely, rejuvenation may arise either from a burial by protecting sediments that precluded TCN accumulation after the deglaciation or from a post-glacial erosion that removed part of TCN concentration while TCN age modelling is based on the assumption that post-glacial denudation rates are equal to zero (Zreda et al., 1994; Putkonen and Swanson, 2003; Putkonen and O'Neal, 2006; Heyman et al., 2011).

Recent advances in TCN dating of alluvium have made it possible to develop reliable fluvial chronologies. Most published ages have been obtained from vertical sediment profiles (Rixhon et al., this issue). This approach provides estimates of the duration of post-depositional sediment exposure and of the erosion rate at the top of the terrace tread abandonment. Reliable ages are produced only if TCN concentrations at the base of the sediment profile have not reached a steady state. The latter is typically observed when high erosion rate affected the top of the profile (sedimentary cover and/or upper part of the fluvial sediments). In that case, the calculated exposure duration corresponds to the minimum age for terrace abandonment. Given that steady state TCN signals are rapidly reached when denudation rates are high, sediment samples should be taken from the centre of the terrace tread. Where profiles are located at the margins of the palaeo-valley, it is difficult to distinguish between TCN signals influenced by vertical (which are related to the nuclide accumulation model in the crust) and lateral (associated with talweg incision) mechanisms. The accuracy of the age also depends on the number of selected samples and the total thickness of the profile, which should ideally exceed 3 m (Rixhon et al., this issue).

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2.2.2. Optically Stimulated Luminescence (OSL) dating

Optically Stimulated Luminescence dating of quartz and feldspar grains has been increasingly used to date the timing of glacial and fluvial sediment deposition (e.g. Straffin et al., 1999; Colls et al., 2001; Peña et al., 2004), and to correlate between glacial and fluvial settings (Lewis et al., 2009). Unlike mineral grains transported through aeolian processes, sediments in glacial and fluvial settings can experience shorter transportation pathways with insufficient sunlight exposure. In fluvial settings, this can lead to incomplete bleaching of the grains, partial resetting of the luminescence signal, and overestimation of the exposure age (Thrasher et al., 2009; Smedley et al., 2016). In a same way, glacigenic sediments can be produced, transported, and deposited subglacially, such that they do not experience any sunlight exposure. A clear understanding of sediment transportation pathways is therefore key for targeted field sampling and accurate interpretation of the OSL signal (see review by Fuchs and Owen, 2008). Evidence suggests that sand sized quartz grains are more suitable for OSL dating than finer sediments (Olley et al., 1998; Colls et al.; 2001; Wallinga, 2002), even though smaller grain sizes are typically transported closer to the water surface. Continued development of single grain and single aliquot (SAR) techniques, mean that is it now possible to assess sediment bleaching characteristics. Although quartz grains are typically preferred, advances in single grain dating of glaciofluvial feldspar (e.g. Smedley et al., 2016), provides further opportunities to develop reliable glaciofluvial chronologies.

2.2.3. Uranium series dating

In carbonate-rich catchments, uranium-series dating has been widely used to date glacial (Hughes et al., 2006, 2010) and fluvial sequences (Candy et al., 2005; Woodward et al., 2008; Adamson et al., 2014). U-series ages date the formation of secondary calcite crystals, such as pedogenic or groundwater calcretes, within a sedimentary sequence. They provide minimum ages of sediment deposition and land surface stabilisation, and have been successfully applied to correlate glacial and fluvial sedimentary sequences in the eastern Mediterranean (Hughes et al., 2006; Woodward et al., 2008; Adamson et al., 2014). Calcite formation largely, though not exclusively, occurs during interglacial periods (e.g. Candy et al. 2005, 2012; Woodward et al., 2008; Adamson et al., 2014). This means that a Pleistocene sedimentary sequence may contain multiple calcite horizons formed during different climate phases (Adamson et al., 2015).As discussed above, accurate interpretation of U-series ages therefore relies on a detailed understanding of the morphosedimentary and topographic context, as well as a detailed and systematic dating programme (Candy et al., 2004).

2.2.4. From dating to age interpretation

Over the last few decades, improvements in Quaternary dating techniques have provided opportunities to examine glacial and fluvial dynamics in more details than was previously possible. Geochronology itself now represents a key tool to assess how river systems have responded to glacial activity. At the Pleistocene timescale, synchronicity between two events (for example, glacial retreat evidenced by the TCN dating of a deglaciated bedrock surface, and fluvial sediments deposited downstream) might imply a causal relationship between them. However, several limitations should be stressed. First, all dating methods are associated with an error range. These ranges have been reduced somewhat by recent methodological advances, but they still often exceed 5 to 10%, which can represent a period of several millennia for sediments deposited during the penultimate glaciations (MIS 6) and older (Early to Mid-Pleistocene). Such limited precision means that sometimes ages cannot be used to securely correlate periods of glacier retreat and fluvial sedimentation/erosion. Second, the resolution of Quaternary dating methods cannot always firmly establish if two events occurred simultaneously. This means that relatively short-term events (less than several millennia), which are preserved in the sedimentary record are not always apparent in the geochronology, even if they indicate a more complex depositional history than a linear fluvial response to glacial change (Ritter and Ten Brink, 1986; Cossart, 2008). This is especially the case for low-frequency-high-magnitude events that occurred during the Pleistocene but which cannot be securely dated. Finally, it has been widely demonstrated that, in the same way that synchronicity does not equate to causality (Vandenberghe, 2012), causality does not always indicate synchronicity, especially when considering paraglacial processes. Research in the Rocky Mountains (Jackson *et al.*, 1982; Church and Slaymacker, 1989; Jordan and Slaymaker, 1991) demonstrates that Holocene to present day fluvial dynamics are largely driven by paraglacial processes that followed the last glaciation, with a response time estimated to 2-10 ka. These limitations show that Quaternary geochronologies must be used as part of a multi-proxy approach, where glacial and/or fluvial deposits are investigated within their depositional and sedimentological context.

2.3 Process based approaches and modelling

Considering the limitations of numerical dating methods, process-based and modelling approaches are also used to examine the relationships between glacial and fluvial dynamics especially in the case of glacially-fed rivers. Process-based analyses can be used to reconstruct meltwater and sediment flux, and hence fluvial response to changes in glacier mass balance.

This approach is based on two conceptual models: 1) the sediment supply model (where an increase or decrease of the fluvial sediment load is associated with accumulation and erosion, respectively); and 2) the stream power model which reflects the capacity of a river to incise. These two conceptual models represent resisting and driving forces in the fluvial system, and should be considered as complementary to allow a reliable reconstruction of the fluvial system evolution. However, they must be used cautiously to avoid over-interpretation (Hanson et al., 2006). This is because they both require high energy, incision and sediment deposition to occur simultaneously (such synchronicity is unlikely, as shown by the research performed within the framework of FLAG). Investigations performed in the Moselle catchment (Cordier et al., 2004, 2006, 2014) underline the need to consider field observations and process-based evidence together, to ensure accurate interpretation.

Modeling of glacial-fluvial system coupling remains a challenging task as it requires both the modeling of the glacial (including the sediment production and transfer) and fluvial system components. This explains why studies focusing on glacial and proglacial areas remain uncommon (De Winter et al., 2012), and are often based on a combination of several, disparate, models that each resemble different parts of the glacial and fluvial systems from source to sink (e.g. glacier flow, glacial erosion, and sediment transport; Kessler et al., 2006, 2008) following a source-to-sink approach.

Combining modeling approaches in this way means that the complexity of glacial-to-fluvial sediment transfer cannot be fully captured (Cossart, 2014). Sediments stored within depocentres during paraglacial phases, for example, interrupt the downstream sediment cascade and are often not represented in such models (Ballantyne, 2003; De Winter, 2012). Further research is therefore necessary to more reliably model distal fluvial response to changes in glacial dynamics.

3. 'Downstream' control on fluvial system: a glacially-induced disruption

During Pleistocene cold periods, downstream glaciation was common in northern European river basins draining towards the North Sea or the Baltic Sea, such as the Thames and the Trent in the UK (White et al., 2010), the Rhine, Elbe, Vistula on the European mainland (Busschers et al., 2007; Starkel, 2003), or in northern Siberia with the main rivers flowing towards the Arctic ocean, as the Ob or Yenissei. It was less common in North America, because the majority of rivers drained south, and therefore away from the major ice masses, since the Tertiary. However, some rivers draining towards the North Atlantic have been influenced by ice cover in their lower reaches, such as the Saint Laurent, the Red or Souris Rivers flowing between the Central United States and Canada (Occhietti, 1990; Bentley et al., 2016). The presence of ice in the ‘downstream’ zone of a pre-existing fluvial system disrupts it in different ways : 1) hydrographical, by influencing river course and valley orientation; 2) hydrosedimentary, by creating new conditions for sediments transport and deposition; and 3) isostatic, the ice pressure generating vertical motions that are particularly significant during glacial and deglacial periods. Hydrographical changes associated to the Pleistocene ice sheets have been recognized in many fluvial systems. Review of studies focusing on european rivers shows that these changes largely depends on the orientation of the pre-existing rivers. In Russia, the large N-S systems such as the Dnieper or the Don were only glaciated in their upper catchments (Starkel et al., 2015), and the glacial influences was rather expressed as ‘upstream’ control –even if this topic was not really considered in past research. It is worth noting that the Scandinavian ice sheet was able to create subglacial depressions, which remained after the ice retreat and were also subsequently used by rivers (Matoshko, 2004). Further west, several mid-european rivers such as the Vistula or the Elbe drain the Hercynian mountains northwards, and are characterized by a S-N orientation (Dvareckas, 2000; Starkel et al., 2007). The presence of the Scandinavian ice sheets hence prevent them from reaching the Baltic or the North Sea during the Pleistocene cold periods, while their progressive retreat northwards allowed the re-establishment of this pathway. Alltogether, this leads to the development of 10-20 km large, typically oriented SW-NE to E-W, ice marginal valleys termed Urstromtäler, ‘fluvial palaeovalleys’ in German (Keihack, 1898; Liedtke, 1981; Liedtke and Marcinek, 2002; Marks, 2004). In the UK, the predominant W-E component of the drainage system explains that the river course was affected or not by the ice-sheet, depending of wether it reaches the valley or not. During the major glacial phase of MIS 12 (the Anglian), the British-Irish Ice Sheet caused the Bytham valley to be diverted southwards towards what is now the present-day Thames valley (Whiteman and Rose, 1992). Similar evolution have been observed in North America, as shown by the formation, during the 1.5-2.4 Ma glaciation, of the modern Ohio valley replacing the buried Teays palaeo-valley (Granger et al., 2001). (Parent, 1987; Granger et al., 2001). The same processes also operate at the local scale inhigh altitude, for example in the Alpine valleys as shown by the capture of the upper Ybbs river in Austria during the penultimate glaciation (Bickel et al., 2015). Changes in river courses could obviously not be an immediate response to ice damming. The formation of glacial lakes is also a common feature associated with the glacially-disrupted rivers. These lakes covered large areas in Asia, where they were fed by Siberian rivers and drained towards the Mediterranean Sea through the Aral and Caspian Seas (Letolle and

Mainguet, 1993; Ehlers, 1996). Similar features were observed in America, where the present-day Great Lakes are inherited from past major lakes (Parent, 1987; Occhietti et al., 2016) and in Europe, such as for example in the Trent catchment (White et al., 2016), in the Warsaw basin in the Vistula catchment (Marks, 2004; Starkel et al, 2007) or in Lithuania (Dvareckas, 2000). Proglacial lakes typically develop during the glacial maxima (Parent, 1987), but lakes may also form during glacial advances (example of the Scarborough lake in Ontario formed during the MIS 5d; Occhietti et al., 2016) deglaciation (Arbogast et al., 2008) or even later, as a result of glacio-isostatic uplift (see below). These lakes are often ephemerial: research performed in the Saint-Laurent area provided evidences for lakes existing during c. 1000 years, as the Vérenderye lake (Occhietti and Richard, 2003; Occhietti et al., 2016).

Ice damming can cause sediment trapping both in these lakes and in the upper valley reaches, and transformation of downstream river flow regime. This is especially the case during deglaciation, as the meltwater from retreating ice sheet allow an increasing of the river discharge. The wide palaeochannels recognized in several mid-european valleys were also attributed to the deglacial period. However, recent chronological studies (Panin et al., 2015; Starkel et al., 2015) demonstrated on the basis of numerical dating that these palaeomeanders formed after the deglaciation –e.g. during the early Holocene- and so were not related to meltwater. In lacustrine palaeoenvironments, fluvial activity may be evidenced by erosional surfaces within lacustrine sediments (as shown for example in the “Don formation” in the Saint-Laurent), or by the recognition of deltaic sediments. The latter may also be used as reliable proxies for lake- or sea-level changes (Parent, 1987; Parent and Occhietti, 1988), related to isostatic adjustement (see below).

Beyond drainage network reorganisation, the presence of glaciers, especially ice sheets, leads to glacio-isostatic adjustment (Bridgland et al., 2010). The magnitude and direction of isostatic change are closely linked to the geography (aerial limit) and timing of ice mass growth and decay. Studies of fluvial systems at the margins of the Scandinavian ice sheet in Western Europe (Busschers et al., 2007), Russia (Panin et al., 2015, this issue), and in North America, demonstrated that areas covered by ice were characterized by subsidence, while the ice periphery was uplifted due to the development of a forebulge. Deglaciation caused isostatic rebound of the formerly glaciated area, but subsidence and disappearance of the forebulge. Fluvial incision driven by isostatic rebound has been recognized in many valleys such as in the Vistula (Starkel et al., 2015; Panin et al., this issue) or in UK and Ireland (Bridgland and Westaway, 2014). In the latter area, a contrast has also been shown between areas glaciated during the MIS 2, where the older terraces have a Lateglacial age, and the non-glaciated areas where older terraces are preserved while the MIS 2-1 fluvial deposits are at the same level as the present-day floodplain: if the presence of older terraces clearly results from the absence of glaciers which would have destructed them (see above), the lack of significant incision since the MIS 2 clearly results from the absence of significant glacio-isostatic rebound. It is finally worth noting that glacio-isostatic adjustement may influence the orientation of the river course, as shown for example in the Dvina valley in Russia (Starkel et al., 2015). Furthermore, the abovementioned lakes may act as local base levels and influence the fluvial response, e.g. by reducing or delaying the post-glacial incision (Dvareckas, 2000). In contrast, significant fluvial incision may be observed as a response of a lake emptying, or of the breaching of an ice-dam (Kasse, 2014; Panin et al., 2015).

Ice masses in the catchment headwaters can also influence fluvial systems through glacio-isostacy. However, unlike downstream glacial activity, these processes are more difficult to identify in high mountain regions, because glaciotectonic signals may be less significant (due to lower thickness of the ice) and because they are superimposed onto other mechanisms of tectonic uplift (see Demoulin et al., this issue), so isolating these signals remains challenging.

3. Glacially fed rivers: unravelling the fluvial response to upstream glacial dynamics

During Pleistocene cold periods, valley glaciers and ice caps developed in many mountain regions including the Pyrenees, Massif Central, Vosges, Alps, Apennines, Dinaric Alps, and Rocky Mountains. These ice masses influenced the behaviour of river systems draining high mountain catchments, as well as those that drained towards lower latitudes, as is the case for the Mississippi in North America, and the Dniepr, Don or Volga in Europe. Upstream glaciation forms the core of this paper, since it corresponds to a situation where glaciers play a key role in driving fluvial dynamics. Unlike downstream glacial activity, the response of river systems to headwater glaciation can be more readily compared to river behaviour in response to climate change in non-glaciated basins. This is because, in a similar way to climatically-driven changes in permafrost or vegetation characteristics (Vandenberghe, 2003, 2008), upstream glaciation can majorly influence catchment hydrology and sediment flux for two reasons: 1) glaciers contribute large volumes of meltwater downstream, and 2) since glaciers are major agents of erosion, they produce vast amounts of sedimentthatare subsequently transported, deposited and reworked by river systems. These glacial controls are largely dependent on the rhythms of Quaternary climate fluctuations, and are manifest in the fluvial system as a succession of climate-sedimentary cycles. These include cold (glacial) periods, cold-to-warm transitions, warm (interglacial) periods, and warm-to-cold transitions (Vandenberghe, 2014). These cycles should be considered together, because the influence of glaciers is not constant through time.

In interglacial periods, mid-latitude glaciers exist only at high altitudes, and their direct influence on river system behaviour is limited to the catchment headwaters. For example, basin-scale fluvial dynamics of the present-day Rhine, Rhone, and Missouri rivers, are not influenced by the presence of glaciers in their source areas. During warm-to-cold transitional periods, as glaciers develop, they begin to store large volumes of freshwater. This storage of water, and its influence on river systems, may be considered similar to the storage associated with the permafrost formation, which is typical in non-glaciated areas subjected to progressive climate continentalization and cooling. In full glacial periods, when glaciers are growing or have stabilized, their influence on catchment water flows is limited, but numerous studies (e.g. Hallet et al., 1996; Koppes et al., 2009) have shown that these periods are associated with the production of large quantities of sediment. This material is transferred to the proglacial floodplain, via meltwater streams and/or mass movement from the valley sides. A strong morphological relationship between frontal moraines and glaciofluvial outwash deposits has been observed in many proglacial areas, and this forms part of the following discussion (Penck and Brückner, 1909; Mandier, 1984; Hein et al., 2009, 2011). Where glacial-fluvial connectivity is high (see discussion below), sediments can be transported and deposited beyond the glacial and proglacial zones, sometimes during a later time period. The influence of glaciers on fluvial sediment load during glacial maxima may be considered similar to that of periglacial slope evolution –bearing in mind that the fluvial dynamics and slope erosion during the coldest periods of the Pleistocene are debated, these periods being associated either with enhanced landscape stability or high morphogenetic activity.

The role of glaciers on meltwater and sediment flux remains significant during deglacial periods (cold-to-warm transitions). This is supported by Holocene and recent glacial and fluvial records, including especially the mediaeval LIA deglaciation. However, this period represents only a short time slice of Quaternary glacial-interglacial cycles, and morphosedimentary records spanning multiple glacial-interglacial cycles are less well preserved. It is important to note that glacier behaviour during deglacial periods is complex. Retreating glaciers can significantly influence river dynamics, but does not always build sizeable morphosedimentary archives, and instead can leave only isolated deposits. Even during cold periods, minor fluctuations of the ice front are not always recorded in the glacial landform assemblage archive, despite major impacts on river systems downstream. Changes in glacier mass balance, regardless of duration or magnitude, can alter river system behaviour by: disturbing the fluvial dynamic (aggradation vs incision); fluvial pattern (channel planform); floodplain geometry (long and cross profiles); and sediment transportation and sorting (Maizels, 1979). This has been shown in the recent evolution of meltwater systems draining present day glaciers, discussed in the following section.

4. Glacially fed rivers: Holocene to present day evidence of a fluvial response to glacial dynamics

Studies of Holocene to modern glacial/deglacial phases, including the post-LGM deglaciation and the Little Ice Age, indicate that proglacial fluvial systems respond rapidly to changes in glacier mass balance. These studies have highlighted three periods of fluvial activity in the geomorphological records, that are used to propose a model of fluvial response to glacier change:

1) Proglacial aggradation while glaciers are growing or have stabilized. This is well validated by research on active proglacial systems at short timescales (10s-100s years; Roussel et al., 2008; Wilkie and Clargue, 2009; Owczarek et al., 2014).

2) Incision in the ice-proximal foreland as glaciers begins to retreat. This is due to the fact that, in their retreat phase, glaciers release large volumes of meltwater while sediment flux remains comparatively unchanged (sediment-limited system; Marren and Toomath, 2013; Owczarek et al., 2014).

3) Paraglacial (e.g. influenced by the evolution of the disappearing glacial system) aggradation at the end of, and after, deglaciation. Church and Ryder (1972, 1989) demonstrated that catchment deglaciation induces a major phase of slope denudation. As glacigenic sediments are released from their temporary storage spaces in the foreland, catchment sediment flux reaches maximal values, and paraglacial aggradation occurs (Jackson et al., 1982; Owen & Sharma 1998; Oetelaar, 2002; Barnard et al., 2004; Barnard et al., 2006). Paraglacial sediment reworking can influence fluvial behaviour for several thousand years after the onset of glacier retreat. The duration and intensity of paraglacial adjustment are linked to 1) the volume of sediment deposited at palaeo-glacier margins, 2) the rate of slope erosion processes, and 3) environmental conditions such as post-glacial climate, timing of vegetation change, catchment size and morphology (Church and Slaymaker, 1989; Harbor and Warburton, 1993; Ballantyne, 2003).

Church and Slaymaker (1989) hence suggested that fluvial response to post-LGM evolution may range from 1 to 10 ka. This means that, in some systems, fluvial behaviour may still be responding to early Holocene glacier retreat. It is possible that more recent phases of glacial activity, such as the Little Ice Age, may have perturbed longer-term trends of paraglacial sediment adjustment. This underlines the complexity of river response to proglacial and paraglacial forcing, but research of this nature is typically limited to short time-scales (a few decades/centuries) and/or small, high altitude or high latitude catchments (Iceland, Svalbard). Furthermore, although it has been shown that paraglacial sedimentation can influence large catchments and sections of valleys far from glaciated areas (Church and Slaymacker, 1989), many empirical studies have focused on glaciofluvial systems in the ice proximal zone, only a few kilometres downstream from the ice front (Roussel et al., 2008; Owczarek et al., 2014). Reconstructions for larger basins and over longer time periods (from the Pleistocene to the Lateglacial) are comparatively limited. They mainly focus either on the uplift associated with deglaciation, as reviewed by Bridgland and Westaway (2014), or on the timing of ice retreat (Böse et al., 2012), and there is little discussion of the downstream fluvial archives.

5. Results: fluvial response to Pleistocene glacial dynamics

Fluvial response to Pleistocene glacial dynamics has been investigated in detail over the last two decades, and there has been a specific focus on European catchments. Key study regions include the mountain catchments of the Alps, the Pyrenees, the Apennines, the Balkans and, at higher latitudes, the Vosges Massif (Figure 1). This section synthetises the existing research in these regions, and explores comparisons between them.

5.1 Alps

The Alps were the largest European massif occupied by glaciers during Pleistocene cold periods – some relic glaciers are still present today. It was also a key field for the development of glacial theory in the 19th Century, and where Penck and Brückner (1909) developed their classic glacial stratigraphic framework in the early 20th Century (Günz, Mindel, Riss and Würm). In fact, their terminology, developed in early Alpine studies was used for several decades by proponents of a climatic origin of fluvial terraces, and applied to many fluvial systems (in both glaciated and non-glaciated areas). Moreover, this led the recognition of four terraces related to the four main glaciations within many fluvial systems, before this "four-glaciations model" was questioned both in the Alps (Billard, 1987) and worldwide, with the development of the isotopic stratigraphy (Shackleton, 1987). Despite this longstanding interest, studies focusing on the

relations between glacial and fluvial dynamics in the Alps remain scarce, apart from a small number of studies in the southern part of the massif.
Recent research in the northern Alps has been focused on the areas of Schaffhausen,in northern Switzerland and in the Ybbs valley in Austria. Quaternary reconstructions for northern Switzerland (Preusser et al., 2011) focused on Pleistocene glacial dynamics, but they suggest that periods of glacial advance were associated with glacio-fluvial sedimentation, while deglaciation (especially at the end of the penultimate glaciation and after the LGM) correspond to profound fluvial incision. This incision was clearly enhanced by the morphostructural conditions (tectonic uplift vs subsidence in the Upper Rhine Graben). Further research into the Quaternary fluvial record and its relationships with the glacial archive in this part of the Alps is required, as the fluvial dynamics are not considered in detail.

The relationships between rivers and glaciers have been investigated in more detailed in a recent study of the Ybbs valley in northern Austria (Bickel et al., 2015). OSL dating demonstrated that, at the end of the penultimate glaciation (MIS 6), glacier retreat led to the deposition of (glacio)fluvial terraces several tens of km downstream from the glacier front. The timing of terrace incision has not yet been established, even if the lower terrace is thought to correspond to the last glaciation; further geochronological investigations are also necessary. In contrast, research in the valleys draining the southern Alps in Italy, especially the Tagliamento, Brenta (Fontana et al., 2008) and the Piave (Carton et al., 2009) provide a high-resolution regional reconstruction of the influence of the last deglacial period on fluvial dynamics. This region is especially valuable because the morphostructural context allows links to be drawn between alpine valleys and the Po-Venetian plain, which is a major sediment depocentre (Fontana et al., 2014).These studies are based on detailed ^{14}C chronologies as well as a small number of luminescence ages from glacial deposits. During the last cold period, glaciers extended to the Po-Venetian plain, facilitating a major phase of downstream sediment transfer, and the formation of large alluvial fans at the contact between the Alpine uplands and the plain. In the Tagliamento valley, the onset of deglaciation at c.18 ka caused incision at the fan apex. Subsequent valley deepening enhanced the transfer of sediments, which accumulate further downstream. The ice distal part of the Brenta valley is also characterized by permanent sedimentation until c.14 ka, although sediment accumulation was much reduced after 18 ka. As glaciers continued to retreat into the mountains, the two major river systems (Tagliamento and Brenta) underwent an incisional phase at c. 14 ka, which continued into the Holocene. The authors show that this incision was first induced by a strong decrease in sediment supply while discharge remained elevated, and subsequently by the fact that the valley and channel long-profiles were no longer in equilibrium with the previous (LGM) gradient. Subsequent aggradation in these systems remains limited for two reasons related to the local context: first, the close proximity of the base level corresponding to the Adriatic Sea, and second, the reduction of glacial-fluvial connectivity, and therefore sediment supply, due to continued glacier retreat.

The neighboring Piave valley (Carton et al., 2009) experienced a somewhat different evolution. This valley is characterized by the presence of an intramontane basin (Vallone Bellunese) separated from the Venetian plain by a gorge valley. Glacier retreat, and catchment deglaciation, from 16-15 ka led to major accumulation of proglacial sediments until c. 8 ka. Further downstream, however, uncoupling of the glacial and fluvial systems, due to glacier retreat, caused a major reduction in sedimentation followed by 'paraglacial-type' incision and the formation of channels into LGM sediments. These channels are sometimes filled with coarse-grained alluvium from a 'paraglacial' accumulation phase, beginning at c. 6 ka. This material is derived from the reworking of material deposited in the intramontane basin during deglaciation. A similar evolution was observed in other catchments (such as the Isonzo) where the occurrence of landslides disconnected the upper and lower parts of the valley, thus preventing significant accumulation in the lowland area between 12 and 7 ka. In some valleys such as the Mincio or Chiese, proglacial lakes have persisted until the present day. The associated moderation of water and sediment flux, explains why these rivers flow in narrow valleys incised into LGM sediments.

5.2 Italian Apennines

Radiocarbon ages of river terraces and alluvial fans in the northern Apennines indicate that major phases of floodplain aggradation occurred during climatic transitional phases (Amarosi et al., 1996). Widespread gravel deposition (19.5-13.0 cal ka BP) corresponds to the onset of deglaciation, when large volumes of meltwater and sediment were delivered downstream. Subsequent incision into the alluvial fill was driven by tectonically-induced base level change. In tectonically-active settings, such as the Italian Peninsula, glacial controls on fluvial dynamics are superimposed onto long-term tectonic characteristics. Accurate identification of glacial drivers in the Quaternary morphosedimentary record must take account of the tectonic context. In the central Apennines, ^{14}C and ^{39}Ar-^{40}Ar ages from glacial and fluvial deposits in the Campo

Felice Basin indicate that major phases of fluvial aggradation correspond to headwater glacial activity during MIS 14, 10, 6, 4, 3, and 2 (Giraudi et al., 2011).

5.3 Pyrenees

The Pyrenean piedmonts have been investigated in detail over the last few years (Figure 2, Table 1), but chronostratigraphical relationships between moraine sequences and river terrace staircases remain unclear. Geochronologies are largely based on ^{14}C, TCN and OSL dating, but there is little systematic cross-dating between glacial and fluvial archives. Regional fluvial correlations are also difficult for three reasons. First, the terrace nomenclature is labelled in ascending order from the valley floors in the northern and eastern parts of the massif, and in the opposite direction in the south. Second, some valleys contain more terrace surfaces than others, preventing reliable correlations from one valley to another. Finally, soil sequences are characterized by leached soils in the northern and eastern Pyrenees, and carbonate soils in the south, making comparisons of relative soil development complicated.

The Quaternary fluvial terraces and frontal moraine deposits of the Pyrenees have been investigated for more than a century (Penck, 1885; Panzer, 1926; Alimen, 1964; Calvet, 2004; Calvet et al., 2011). There is a general consensus that moraine formation corresponds to major phases of fluvial aggradation, during Pleistocene cold periods. The Lannemezan fluvial formation is the highest fluvial surface in the northern piedmont, and overlies a very old (assumed middle Pleistocene or older) glacial till (Hétu et al., 1989, 1992). In the easternmost part of the range, an assumed Lower/Middle Pleistocene terrace is correlated to the Carol frontal moraine, which is assumed to be the oldest of the Pyrenean range (Calvet, 2004; Calvet et al., 2011). In the Ariège valley, terraces T2 and T3 were TCN dated, and correlate to MIS 6 and 8, respectively, while the older moraines were dated by ^{10}Be to the end of MIS 6 (Delmas et al., 2011, 2015). In the southern part of the Pyrenees, a fluvial terrace has been OSL dated to MIS 6 in the Cinca and Gallego valleys, and correlated to moraine deposits with stripped boulders (Peña et al., 2004; Sancho et al., 2003, 2004; Lewis et al., 2009). In the Aragon valley, however, only the outermost frontal moraine yielded an MIS 6 age (OSL age of 171±22 ka), while the high terrace of Castiello de Jaca, which is morphologically correlated to this moraine, might date to MIS 8, since it correlates on the basis of relative elevation and pedological evidences to the MIS 8 fluvial terrace (OSL age of 263±4.8 ka) found in the valley of the Subordan Aragon, a tributary of the Aragon (García Ruiz et al., 2013). Further east, the piedmont fans of the Sègre and Nogueras, partly dated by TCN to MIS 4-7 (Stange et al., 2013), are disconnected from the end moraines. Numerical ages of the older terraces of the Cinca and lower Gallego could not be established, but the four highest levels exhibits a reverse palaeomagnetism which suggests an age older than 780 ka (Benito et al., 1998, 2010). This interpretation has been recently confirmed by an ESR dating at 1276±104 ka and paleomagnetism data on the higher level (+ 160 m) of the Alcanadre river (Sancho et al., 2017). In both cases, dating uncertainties are too large to allow the timing of terrace formation to be accurately correlated with a specific climate phases.

In contrast, data are more precise for the last glacial cycle. On the northern part of the Pyrenees, the Würmian maximum ice extent (MIE) is attributed to MIS 4 on the basis of TCN ages in the Ariège valley and ^{14}C and palynological data from ice marginal and proglacial lake sediments in the Garonne, Gave de Pau, and Gave d'Ossau valleys (Andrieu et al., 1988; Andrieu, 1991; review in Calvet, 2004; Calvet et al., 2011; Delmas, 2015). During the Global LGM (24 and 19 cal. ka BP, MARGO Project, 2009), the ice marginal position in the Ariège valley was c.7 km upstream of the MIS 4 ice extent (Delmas et al., 2011). Three TCN profiles were performed on the lowest terrace (T1) which is topographically linked to the Global LGM terminal moraine. The profiles are located 4, 22 and 53 km downstream of the Global LGM ice terminal position and provide ages of $17.5^{+2}_{-3.5}$ka, $13.8^{+3.6}_{-0.4}$ka and$13^{+3.5}_{-0.5}$ ka, respectively. Using these fluvial TCN ages, as well as 34 dates from glacial boulders and ice scoured bedrock surfaces in the Ariège catchment (Delmas et al., 2011),four phases of terrace T1 development have been identified: 1) A major phase of aggradation occurred when the Ariège trunk glacier reached the north Pyrenean foreland. No terrace level corresponding to MIS 4-3 has been identified between T1 and T2, and it is assumed that this proglacial aggradational phase lasted from the MIS 4 to the Global LGM. 2) A short incision phase occurred at the end of the Global LGM, when ice retreated into the upper part of the catchment. Incision was confined to the ice proximal zone, as shown by two surfaces inset into terrace T1, and observed only until 10km downstream from the Global LGM ice terminus; 3) Another period of aggradation occurred from the end of the LGM until the end of the Bølling/Allerød or the early Holocene. At that time, the Ariège catchment was almost completely deglaciated, and aggradation was instead driven by paraglacial adjustment; 4) Regional palynological data indicate that the Bølling/Allerød corresponds to the first phase of vegetation recolonisation in the Pyrenees. The upper limit of the treelinewas located at 1800 m asl, and extended to 2000 m at the Lateglacial-Holocene transition (Reille and Andrieu, 1993). This suggests that the incision period that marked the end of paraglacial sedimentation is likely to be a consequence of decreased sediment input from the slopes (and so increased stream power), due to enhanced vegetation cover.

In the Garonne, Gave de Pau, and Gave d'Ossau valleys, LGM ice extent is less precisely delineated than the MIS 4 margins. Hence, comparisons between Würmian glacier fluctuations and glaciofluvial activity and terrace formation (T1) is more complex than that in the Ariège valley. However, recent TCN data from the Aspe and Garonne valleys (Nivière et al., 2016) suggest that valley evolution followed a similar pattern to that of the Ariège valley. In the Aspe Gave valley, the incision of terrace T1 was dated to 18±2 ka in the ice proximal zone. In the Garonne valley, the lower terrace is dated to $14.6^{+9.6}_{-4.3}$ka at the foot of the Würmian MIE terminal moraines and to $13.1^{+6.7}_{-3.9}$ka 40 km further downstream (Stange et al., 2014).

In the Southern part of the massif, the chronology suggests a much more complex evolution, with several stepped terraces attributed to the last glacial cycle. An MIS 5 terrace (c. 100 ka) has been dated in the Cinca (OSL dating; Lewis et al., 2009) and the Segre (TCN dating; Stange et al., 2013) valleys, but there are no ages available for the glacial deposits, except for two inconsistent OSL ages from Aurin in the Gallego catchment (85 and 38 ka). Another major terrace was dated to MIS 4 in the Segre (TCN), the Cinca, Gallego and Aragon (OSL) valleys. A contemporaneous till was dated by OSL in the Cinca, where it corresponds to the Würmian maximum ice extent (MIE), and in the Aragon (innermost moraine of Castiello de Jaca). No similar evidence was found in the Gallego valley, and the age of the Aurin moraine remains hypothetical (see above). In the Cinca valley, an MIS 3 terrace has been identified on the basis of OSL ages, but these are characterized by high scattering (mean age 51±4 ka). A similar scattering was observed in the Gallego valley (OSL ages of fluvial sediments range from 55 to 32 ka), and correlation with the Senegüe moraine (dated to 36±2 and 36±3ka by OSL) remains uncertain (Lewis et al., 2009; Benito et al., 2010; García Ruiz et al., 2013). In the Valira valley in Andorra, a till sequence overlying an alluvial fan has been OSL dated to 32.7±1.1ka (Turu et al., 2016). In the southwest Pyrenees, the spatial extent of the Cinca and Gallego trunk glaciers during the Global LGM is not well known. It is likely that this is due to palaeoclimatic reasons such as aridity and weak westerly winds. However, a low altitude terrace surface has been identified in these valleys. It is less well preserved than the higher terraces observed elsewhere, and yields OSL and ^{14}C ages of 22-9 ka. Accordingly, terrace incision is correlated to the Lateglacial-Holocene transition, which is consistent with observations in the northern part of the massif (Lewis et al., 2009; Benito et al., 2010).Fluvial terraces are also correlated with Heinrich events, where aggradation is associated with increased meltwater discharge. The limited extent of the MIS 2 fluvial terrace could also be related to the lack of significant water supply at that time.

5.4 Eastern Mediterranean

The mountains of the Balkans were also glaciated at multiple occasions during the cold stages of the Pleistocene. There is an increasing body of research into the Pleistocene glaciofluvial record of this region (e.g. Woodward et al., 2008; Adamson et al., 2014; 2016a, b; Žebre and Stepišnik, 2015). U-series and electron spin resonance (ESR) dating of alluvial records from the limestone-dominated Voidomatis basin, northwest Greece, show high sedimentation rates during MIS 5d-2, but Middle Pleistocene fluvial deposits are not well-preserved (Woodward et al., 2008). This contrasts with the evidence of headwater glacial activity, which shows that glacier extent during the last cold stage was limited compared to the major glacial advances of MIS 12 and 6 (Hughes et al., 2006). The late Pleistocene alluvial record may reflect a cumulative signal of glaciofluvial sediment delivered downstream and reworked over multiple glacial cycles. The Voidomatis record contrasts with the Pleistocene glaciofluvial deposits of the Orjen massif in Western Montenegro, where thick deposits of alluvium from MIS 12 are well-preserved. U-series ages and sedimentology indicates that the majority of the sediment was deposited during a single depositional phase. Sediment corresponding to more recent glacial phases, during MIS 6 and 5d-2, are either absent or present as only thin veneers on top of the Middle Pleistocene deposits. It must be remembered that U-series methods provide minimum ages of sediment deposition. It is likely that maximum alluviation occurred at the end of the glacial phases, but this cannot be resolved using U-series techniques. Unlike the Voidomatis record, the alluvial sequences at Orjen reflect major changes in glacial-fluvial system coupling since the Middle Pleistocene. During MIS 12, glacier margins advanced from the massif into the surrounding basins, and large volumes of sediment were deposited in poljes, steep sided river valleys, and as alluvial fans. One of the largest alluvial fans developed at thesouthern margin of the Orjen massif, and has since been partially submerged by rising sea level (Adamson et al., 2016b). During subsequent glacial phases, glaciers did not extend beyond the massif, and large areas of limestone bedrock were exposed. Meltwater and sediment were channelled into the subterranean karst, effectively decoupling the glacial and fluvial systems (Adamson et al., 2014; Žebre and Stepišnik, 2015). Since MIS 12, there has been very little incision into the depocentres surrounding Orjen, and the alluvial fill is extremely well-preserved.

Despite research in Eastern Mediterranean does not provide a high-resolution reconstruction of the fluvial response to glacial dynamics, it highlight the importance of the glacial-fluvial system coupling and the way this coupling is influenced by hydrogeology and topography, and how this may explain the formation and preservation of the sedimentary record.

5.5 Vosges massif and its surroundings

Despite its relatively small extent and low altitude (less than 1500 m asl), a good deal of research has focused on the regional glacial history and glaciofluvial dynamics of the Vosges Massif (Seret, 1966; Seret et al., 1990; Flageollet, 2002) and on fluvial dynamics of rivers draining formerly glaciated areas. The main rivers draining the Vosges Massif are the Moselle, Meurthe and Sarre, belonging to the Rhine catchment, and the Ognon flowing towards the Saône and the Rhône (Figure 3; Cordier et al., 2006, 2012, 2014 ; Madritsch et al., 2012). Despite intensive investigation and high resolution mapping, no obvious morphological continuity could be proposed to firmly correlate the glacial, glacio-fluvial and fluvial landforms and deposits. This may first be explained by the presence of a gorge section (Moselle valley in the horsts of Epinal, Figure 3) and/or large morphostructural depressions (Moselle valley near Remiremont, Meurthe valley near Saint-Dié, Figure 3) which alter glacial-fluvial connectivity. This may also be explained by the fact that many evidences of Over the last decade, detailed sedimentology coupled with OSL dating, has made it possible to unravel the influence of deglacial periods on the Moselle and Meurthe rivers. Fluvial terraces of the Meurthe valley downstream from the Vosges Massif contain a thick lower unit mainly composed of sandy sediments coming from the non-glaciated areas (Cordier et al., 2006; Occhietti et al., 2012). This unit is locally characterized by the presence of cryoturbation features. It is eroded in its upper part, and overlain by coarser sediments with a high proportion of granite coming from the glaciated part of the massif (Figure 4). A similar sequence has been described in the Moselle valley (Cordier et al., 2014), although grain-size and petrographic contrasts between the lower and upper deposits are less pronounced. OSL ages indicate that the lower unit (cold-period deposits) has been considerably reworked, and the following reconstruction has been proposed: the release of meltwater during early deglaciation promotes significant erosion in the fluvial system. This is especially apparent in deposits from older, Pleistocene deglacial phases, both in ice proximal areas (a few tens of km away from the glacier front) and further downstream. This was also enhanced by the trapping of the sediment in proglacial lakes formed during deglaciation (especially after the LGM; Flageollet, 2002). Lateral fluvial erosion was dominant, but there is evidence for localized vertical incision down to bedrock, especially in the axis of the palaeovalleys. This deepening, however, does not result in the abandonment of the terrace. The relative weakness of vertical incision may be due to sustained sediment load, for example, due to slope erosion (especially in the lower Moselle valley flowing through the Rhenish Massif) or from the reworking of sediments deposited in the valleys during the cold period - before the release of the glacial load from the Vosges Massif. It may occur in conjunction with the persistence of a braided channel patterns linked to high energy conditions (high discharge and load). The concentration of water in a single channel occurs only when returning to interglacial conditions.

6. Discussion

6.1 Fluvial response to glacial fluctuations

The catchments analysed in this study indicate that fluvial aggradation dominantly occurs during two main periods: glacial advance, when ice masses are actively eroding and exporting sediment downstream; and deglaciation, when meltwater flux is increased and can mobilise large volumes of glacigenic sediments (Figure 5). Deglacial phases (cold to warm climate transition) are periods of major landscape evolution. Sediments become exposed by a receding ice margin and valley slopes are not yet stabilised by vegetation. This presents a vast source of readily erodible material that can be entrained, transported and deposited downstream, until sediment supply becomes exhausted. Incision into the alluvial fill is associated with sustained high meltwater discharge conditions coupled with lower sediment yields. These conditions have been identified at the onset of the deglacial phase ("deglacial 1" in Figure 5) and/or towards the end ("deglacial 3"), the latter being related to increased vegetation density. It is worth noting that this incision may also be influenced by glacioisostasy (Occhietti et al., 2016). However, further research are required to validate this assumption, as the extent of the areas affected by such an isostatic rebound are generally not known in Europe.

In some settings, paraglacial slope denudation and remobilisation of pre-existing glacial/glaciofluvial sediment has caused renewed aggradation ("deglacial 2", Figure 5). This is especially evident in the Alps, the Italian Apennines and the Ariege basin of the Pyrenees, where catchments are still responding to Holocene deglaciation (Delmas et al., 2015). It has also been seen in the Voidomatis basin, Greece (Woodward et al., 2008), where sediment from the last cold stage bears the sedimentary signature of glacigenic material delivered to the basin during previous glacial phases in MIS 12 and 6. In the Mediterranean, vegetation can quickly recolonise and stabilise a deglaciating catchment, and the paraglacial period is short-lived. This contrasts with Alpine catchments, where land surface stabilisation is more prolonged and the paraglacial 'window' is much extended. In other basins, such as the depocentres surrounding Orjen, Western Montenegro, there is no significant evidence of paraglacial sediment reworking, and meltwater and sediment dynamics are strongly controlled by catchment topography and hydrogeology (Adamson et al., 2015, 2016a). It is only through detailed sedimentology, and geochronological analysis, such as OSL, U-series, and TCN, that primary depositional phases can be distinguished from long-term paraglacial sediment dynamics. This is especially effective for the last deglaciation, but may also be assumed for older glacial periods : focusing on the upper Dnieper, Matoshko (2004) hence suggests that aggradation took place during the post-MIS 8 deglacial period. This assumption must, however, be confirmed : reconstructions are actually more challenging for old archives (e.g. Mid-Pleistocene and younger), if sedimentary sequences represent a palimpsest of multiple aggradation and reworking phases. This is because the uncertainties associated with Quaternary dating methods increase with sediment age, so that even if the age can be constrained to an individual deglacial phase, the dating uncertainty can be too high to unravel whether the sediments were deposited directly by meltwater, or several thousands of years later when glacier activity was negligible (Cordier et al., 2014). The key issue here is not so much the timing of sediment creation (e.g. rock erosion in relation to glacial or paraglacial processes), but the timing of sediment transport, which directly relates to the connectivity between glacial and fluvial systems.

6.2 The role of glacial-fluvial connectivity

Glacial-fluvial system connectivity is important in the production and preservation of the morphosedimentary record (Figure 5). Considering an individual glacial-interglacial cycle, if glacial and fluvial systems are well-coupled, meltwater and sediment are delivered directly downstream, and their records can be securely correlated. In the Colorado Front Range of the Rocky Mountains, Schildgen et al. (2002) associate fluvial aggradation with deglaciation phases ("deglacial 2", Figure 5), when the meltwater is able to transport large quantities of glacial sediments. They conclude that TCN dating of fluvial terraces may even provide a reliable marker for glacier retreat. In contrast, where proglacial lakes, intramontane basins, karst terrain, and alluvial fans interrupt the meltwater and sediment cascade, fluvial systems might not be responding directly to glacial activity. As evidenced by glaciated basins in the Southern Alps (Fontana et al., 2014), proglacial lakes can store and release sediments independently of primary glacial erosion and meltwater transport. With progressive glacier retreat, the glacial and fluvial systems can become increasingly decoupled, and local topographic conditions control the nature of the fluvial archive (e.g. Carton et al., 2009; Madritsch et al., 2012). The Combe d'Ain glaciolacustrine complex in the Jura is associated with prograding deltaic sediments and glacial deposits, with evidence for fluvial erosion during deglaciation. The lacustrine sequences indicate that this erosion strongly depends on base-level change (Kasse, 2014) and the fluvial system power (Campy, 1982; Passmore and Waddington, 2009).

Over multiple glacial-interglacial cycles, changes in glacial and fluvial system coupling have major impacts on the morphosedimentary record. In the karst terrain of western Montenegro, meltwater and sediment were increasingly channelled into the subterranean karst networks after the major glaciation of MIS 12. These hydrogeological controls on meltwater and sediment routing, as well as cementation by secondary carbonates, have protected the Middle Pleistocene (MIS 12) records from subsequent incision and reworking. Surficial evidence from more recent depositional phases (MIS 6 and 5d-2) is limited (Adamson et al., 2014). This contrasts with other European fluvial archives, where the oldest Pleistocene deposits have been reworked and sediments from more recent glacial phases are well-preserved (e.g. Woodward et al., 2008, Lewis et al., 2009).

In addition to sediment interception by intramontane basins and karst terrain, alluvial fans often develop in the glacial-fluvial transitional zone, especially downstream of confined valley sections. They can contain large volumes of sediment that can profoundly alter the morphology of the transitional area. In the Moselle valley, the well-preserved Noirgueux fan complex is associated with the frontal moraine of the last glaciation (Flageollet, 2002) as well as a suite of fluvial terraces downstream of the moraine, and several lacustrine terraces preserved upstream. Similar fans have been recognized further downstream in the Moselle valley, north of Epinal (Harmand and Cordier, 2012). They can be morphologically correlated to older glaciations (Flageollet, 1988), but no age control is available. However, this shows that successive glaciations can produce similar glaciofluvial landforms preserved along the valley, in relation to the former ice-marginal position.

6.3. Ice proximal versus ice distal fluvial response?

It is commonly assumed, in catchments that were only glaciated in their headwaters, that the influence of glacial activity decreases with distance downstream. Establishing spatial changes in the relative impacts of glacial processes is key for accurate interpretation of the fluvial record. This is especially important in large basins, where river systems are many kilometres long, and may be fed by tributaries delivering both glacial and non-glacial sediments. This is the case for the Lower Garonne (SW France), which is fed by rivers draining the Massif Central, and for the Moselle which, in its lower course, flows through the Rhenish Massif and contains fluvial terraces composed of gravels from the glaciated part of the Vosges Massif. With increasing distance downstream, the impacts of glaciation may become negligible where local sediment input is high and/or if glacial sediments from the catchment headwaters are trapped and stored along the valley, in landforms and proglacial lakes for example. The influence on the water discharge is similarly reduced, due to the increasing size of the catchment in the downstream part of the valley and hence to the increasing contribution of periglacial tributaries. In fluvial systems flowing parallel to an ice margin (as was the case for the Trent; White et al., 2010; Bridgland and Westaway, 2014) the decreasing effect of the glacial system with increased distance downstream is less obvious, because the glacially-fed tributaries are able to influence the evolution of the whole fluvial system.

The studies explored here indicate that the influence of glaciers does not change linearly with increasing distance from the ice front. In the Italian Alps, alluvial records highlight the complexity of river response to deglaciation at the end of the Pleistocene: a first phase of fluvial activity is associated with ice proximal aggradation and distal erosion; a second phase is associated with stability in the ice proximal area and distal aggradation. The morphostructural conditions of the valley were found to be as important as the distance from ice margins in conditioning fluvial response to deglaciation. In the Moselle catchment, research underline that a main period of sediment reworking took place at the end of the Saalian. Evidence for this reworking was found along the whole valley from the vicinity of the Vosges Massif to the Paris Basin and the Rhenish Massif (Cordier et al., 2014). The imprint of deglaciation is clear in the upstream part of the valley, while other processes associated to the periglacial conditions (melting of the snow or the permafrost) must be considered to explain the increased discharge allowing erosion in the downstream course. This indicates that fluvial evolution of a glaciated valley can be driven not only by glacial dynamics, but also periglacial and non-glacial processes.

6.4. Internal (glacial) versus external (climate and tectonic) forcing mechanisms

Because glacial and periglacial processes are driven by climate change, the impacts of these processes on river system behaviour should be considered as part of 'climate forcing' as defined by Büdel (1977) and updated within the framework of the Fluvial Archives Group (e.g. Vandenberghe, 2003, 2008, 2014; Bridgland and Westaway, 2007). A key question is whether

glacially-fed rivers exhibit a specific behaviour when compared to non-glacially fed rivers of similar size, lithology, tectonics, or base level (provided that various conditions may be active simultaneously and occuring in superposition to each other in a given catchement).
Recent analysis of Quaternary morphosedimentary records in North American catchments deal with this comparison. Hanson et al. (2006) focus on two catchments in the Eastern Rocky Mountains: the Laramie River, which was partly glaciated during Pleistocene cold periods; and its tributary Sybille Creek, which was not glaciated. A combination of field investigation, OSL dating, and process-based analyses, indicated that both catchments experienced a similar evolution pattern regardless of the presence of glaciers. However, it is worth noting that the Laramie catchment is five times larger than the Sybille Creek catchment, andthe study area lies at the confluence between both rivers, >100 km downstream of the glaciated part of the Laramie catchment.

In the Western Rocky Mountains, California, Dühnforth et al. (2008) have examined Late Pleistocene sediment dynamics in neighbouring catchments. Alluvial fan sequences indicate that catchments with extensive glacier cover were characterised by high sediment flux and high amplitude fluctuations between aggradation and incision. Incisional phases were triggered by sediment trapping in the glaciated part of the catchment. In contrast, variations in sediment load in non-glaciated catchments were less pronounced, and a more regular sediment throughput preventd intensive incisional phases.

In Europe, evidence for a specific fluvial response to glacial activity (in comparison to non-glaciated rivers) has been identified in the Eastern paris Basin for the rivers draining the Vosges Massif (Cordier et al., 2012, 2014). Morphological, sedimentological, and geochronological investigations indicate that a significant incisional period (>12 m) occurred in the upper valley of the Sarre near Sarrebourg (Figure 3) at the end of the Saalian. The Sarre catchment remained more or less ice-free during Pleistocene cold periods. This contrasts with the neighbouring Moselle and Meurthe valleys, where vertical erosion was much less pronounced (a few metres), and instead lateral erosion and reworking of cold-period sediments affected the whole system. The most plausible explanation, derived from the available geochronological framework, is therefore that incision in the Sarre valley at the end of the Saalian, was a product of enhanced streamflow due to snowmelt. This explanation is consistent with the morphoclimatic context of the area. Incision in the Sarre valley must also be attributed to the fact that the removal of sediments deposited under periglacial conditions during the previous cold period was not followed by an increase in sediment load as was the case in the neighboring Moselle and Meurthe valleys. This lack of sediments results in our view from a combination between 1) a reduced sediment input from the headwaters (in relation with the absence of developed glacial system) and 2) a small contribution of the proximal areas, clearly underlined by the sediment lithology (predominance of siliceous deposits from the Vosges Massif, while most of the upper catchment is developed in the limestones and marls of the Eastern Paris Basin; Harmand, 2007).

The Pleistocene alluvial records synthesised in this review demonstrate that glacial activity can profoundly modify fluvial behaviour, even if the impacts are constrained to small or locally glaciated catchments, or where glacial-fluvial connectivity reduces the direct role of glaciers. The inherent relationship between glacier dynamics and climate, means that fluvial response to glaciation must be also considered in a climatic context. In a similar way, fluvial behaviour is also superimposed onto tectonic and base-level changes. In glaciated basins, fluvial incision, for example, should not only be related to glacier behaviour, but also to the wider context of tectonic uplift – which is typical of glaciated mountains regions. Research in Italian river basins demonstrates the importance of base-level change (namely post-glacial sea level rise) in determining sedimentation pattern. Aggradation is dominant in the coastal (piedmont) plain.In the Italian Alps and Apennines, rates of incision are strongly conditioned by base level change and tectonic uplift (e.g. Amorosi et al., 1996, Fontana et al., 2008) and glacial controls on river dynamics are superimposed onto this regional tectonic framework.

7. Conclusion and perspectives

This first review paper dedicated to fluvial response to glacial dynamics underlines the complexity of the interactions between glacial and fluvial systems, and the importance of the meltwater and sediment coupling. Using research from various European and Northern American catchments, we propose a general scheme of evolution for rivers affected by the presence of glaciers in their headwaters, which includes both erosional and aggradational patterns. Further research is, however, required to improve this model especially by improving the temporal resolution (except for the last glacial period which is relatively well constrained) and by providing a better insight on the spatial variability of the fluvial response, depending on the various parameters that were highlighted in this study (proportion of the catchment being glaciated, location in the catchment, morphological context etc.). Further investigations are also required to unravel the influence, in addition of the external forcing, of the internal control, in particular to explain the incision observed during glacier retreat and observed in various fluvial systems during the Pleistocene (Bridgland and Westaway, 2014) or currently for example in Iceland or Spitsbergen.

This study highlights the ability of the Fluvial Archives Group to promote original research topics and to investigate them by associating field-based approach, modern techniques (geochronology and modelling), and by including comparison between different study areas, which is key for our ability to isolate the glacial influence on fluvial systems.

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References

Adamson, K. R., Woodward, J. C. and Hughes, P. D., 2014. Glaciers and rivers: Pleistocene uncoupling in a Mediterranean mountain karst. Quaternary Science Reviews 94, 28-43.

Adamson, K., Candy, I., Whitfield, L., 2015. Coupled micromorphological and stable isotope analysis of Quaternary calcrete development. Quaternary Research 84, 272-286.

Adamson, K. R., Woodward, J. C., Hughes, P. D., 2016a. Middle Pleistocene glacial outwash in poljes of the Dinaric karst. In Gao, Y. and Alexander Jr, E.C (Eds) Caves and Karst Across Time (Vol. 516). Geological Society of America 247-263.

Adamson, K.R., Woodward, J.C., Hughes, P.D., Giglio, F., Del Bianco, F., 2016b. Middle Pleistocene glaciation, alluvial fan development and sea-level changes in the Bay of Kotor, Montenegro. Geological Society, London, Special Publications 433-13.

Alimen, H., 1964. Le Quaternaire des Pyrénées de la Bigorre. Mem. Serv. Carte géol. France, 394 pp.

Amorosi, A., Farina, M., Severi, P., Preti, D., Caporale, L., Di Dio, G., 1996. Genetically related alluvial deposits across active fault zones: an example of alluvial fan-terrace correlation from the upper Quaternary of the southern Po Basin, Italy. Sedimentary Geology 102, 275-295.

Andrieu, V., 1991. Dynamique du paléoenvironnement de la vallée montagnarde de la Garonne (Pyrénées centrales, France) de la fin des temps glaciaires à l'actuel. Thèse de Doctorat de 3e cycle, Université de Toulouse-le-Mirail, 311 pp.

Andrieu, V., Hubschman, J., Jalut, G., Hérail, G., 1988. Chronologie de la déglaciation des Pyrénées françaises. Dynamique de sédimentation et contenu pollinique des paléolacs:

application à l'interprétation du retrait glaciaire. Bulletin de l'Association Française pour l'Etude du Quaternaire 34, 55–67.

Antoine, P., Lautridou, J.-P., Laurent, M., 2000. Long-Term Fluvial archives in NW France: Response of the Seine and Somme Rivers to Tectonic movements, Climatic variations and Sea level changes. Geomorphology, 33, 183-207.

Antoine, P., Limondin-Lozouet, N., Chaussé, C., Lautridou, J.-P., Pastre, J.-F., Auguste, P., Bahain, J.-J., Falgueres, C., &Ghaleb, B., 2007. Pleistocene fluvial terraces from northern France (Seine, Yonne, Somme): synthesis and new results. Quaternary Science Reviews, 26, 22-24, 2701-2723.

Arbogast, A.F., Bookout, J.R., Schrotenboer, B.R., Lansdale, A., Rust, G.L., Bato, V.A., 2008. Post-glacial fluvial response and landform development in the upper Muskegon River valley in North-Central Lower Michigan, U.S.A. Geomorphology 102, 615–623.

Ballantyne, C.K., 2003. Paraglacial landform succession and sediment storage in deglaciated mountain valleys: theory and approaches to calibration, Zeitschrift für Geomorphologie 32, 1-18.

Barnard, P.L., Owen, L.A., Finkel, R.C., 2004. Style and timing of glacial and paraglacial sedimentation in a monsoonal-influenced high Himalayan environment, the upper Bhagirathi Valley, Garhwal Himalaya. Sedimentary Geology 165, 199–221.

Barnard, L., Lewis, A., Finkel, R.C., 2006. Quaternary fans and terraces in the Khumbu Himal south of Mount Everest: their characteristics, age and formation. Journal of the Geological Society 163, 383–399.

Benito, G., Pérez-González, A., Gutiérrez, F., Machado, M.J., 1998. River response to Quaternary large-scale subsidence due to evaporite solution (Gállego River,Ebro Basin, Spain). Geomorphology 22, 243–263.

Benito, G., Sancho, C., Peña, J.L., Machado, M.J., Rhodes, E.J., 2010. Large-scale karst subsidence and accelerated fluvial aggradation during MIS6 in NE Spain: climatic and paleohydrological implications. Quaternary Science Reviews 29, 2694–2704.

Bentley, S.J. Sr., Blum, M.D., Maloney, J., Pond, L., Paulsell, R., 2016. The Mississippi River source-to-sink system : perspectives on tectonic, climatic and anthropogenic influence, Miocene to Anthropocene. Earth-Science Reviews 153, 139-174.

Bickel, L., Lüthgens, C., Lomax, J., Fiebig, M., 2015. Luminescence dating of glaciofluvial deposits linked to the penultimate glaciations in the Eastern Alps. Quaternary International 357, 110-124.

Billard, A., 1987. Analyse critique des stratotypes quaternaires. Editions du CNRS, 143 p.

Boenigk, W., Frechen, M., 2006. The Pliocene and Quaternary fluvial archives of the Rhine system. Quaternary Science Reviews 25, 550–574.

Böse, M., Lüthgens, C., Lee, J.R., Rose, J., 2012. Quaternary glaciations of northern Europe. Quaternary Science Reviews 44, 1-25.

Bridgland, D.R., 2010. The record from British Quaternary river systems within the context of global fluvial archives. Journal of Quaternary Science 25, 433-446.

Bridgland, D.R., Westaway R., 2007. Climatically controlled river terrace staircases: A worldwide Quaternary phenomenon.Geomorphology 98, 285-315.

Bridgland, D.R.,Westaway, R., 2014. Quaternary fluvial archives and landscape evolution: a global synthesis. Proceedings of the geologists' Association 125, 600-629.

Bridgland, D.R., Westaway, R., Howard, A.J., Innes, J.B., Long, A.J., Mitchell, W.A., White, M.J., White, T.S., 2010. The role of glacio-isostasy in the formation of post-glacial river terraces in relation to the MIS 2 ice limit: evidence from northern England. Proceedings of the Geologists' Association 121, 113–127.

Briner, J., Miller, G.H., Davis, P.T., Finkel, R., 2006. Cosmogenic radionuclides from fjord landscapes support differential erosion by overriding ice sheets. Geological Society of America Bulletin 118, 406–420.

Büdel, J., 1977. Klima-Geomorphologie. Gebrüder Bornträger, Berlin, 304 pp.

Busschers, F. S., Kasse, C., van Balen, R. T., Vandenberghe, J., Cohen, K. M., Weerts, H. J. T., Wallinga, J., Johns, C., Cleveringa, P., Bunnik, F. P. M., 2007. Late Pleistocene evolution of the Rhine-Meuse system in the southern North Sea basin: imprints of climate change, sea-level oscillation and glacio-isostacy. Quaternary Science Reviews 26, 3216–3248.

Calvet, M., 2004. The Quaternary glaciation of the Pyrenees. In: Ehlers, J., Gibbard, P. (Eds.), Quaternary Glaciations—Extent and Chronology, Part I: Europe. Developments in Quaternary Science, vol. 2a, Elsevier, Amsterdam, 119–128.

Calvet, M., Delmas, M., Gunnell, Y., Braucher, R., Bourlès, D., 2011. Recent advances in research on Quaternary glaciations in the Pyrenees. In : Ehlers, J., Gibbard, P., Hughes, P., (Eds.), Quaternary Glaciations, Extent and Chronology, a closer look, Part IV, Developments in Quaternary Science, vol. 15, Elsevier, Amsterdam, 127–139.

Campy, M., 1982. Le Quaternaire franc-comtois. Essai chronologique et paléoclimatique. Thèse d'État, Géologie, Université de Besançon, 575 pp.

Candy, I., Black, S. and Sellwood, B.W., 2005. U-series isochron dating of immature and mature calcretes as a basis for constructing Quaternary landform chronologies for the Sorbas basin, southeast Spain. Quaternary Research 64, 100-111.

Candy, I., Black, S. and Sellwood, B.W., 2004. Quantifying time scales of pedogenic calcrete formation using U-series disequilibria. Sedimentary Geology 170, 177-187.

Candy, I., Adamson, K., Gallant, C.E., Whitfield, E., Pope, R., 2012. Oxygen and carbon isotopic composition of Quaternary meteoric carbonates from western and southern Europe: their role in palaoenvironmental reconstruction. Palaeogeography, Palaeoclimatology, Palaeoecology 326, 1-11.

Carney, F., 1907.A form of outwash drift. American Journal of Science 23, 336-341.

Carton, A., Bondesan, A., Fontana, A., Meneghel, M., Miola, A., Mozzi, P., Primon, S., Surian, N., 2009. Geomorphological evolution and sediment transfer in the Piave River system (northeastern Italy) since the Last Glacial Maximum. Géomorphologie : relief, processus, environnement 3, 155-174.

Church, M., Ryder, J.M., 1972. Paraglacial sedimentation, a consideration of fluvial processesconditioned by glaciation.Geological Society of America Bulletin 83, 3059–3072.

Church, M., Ryder, J.M., 1989. Sedimentology and clast fabric of subaerial debris flow facies in a glacially influenced alluvial fan—a discussion. Sedimentary Geology 65, 195–196.

Church, M., Slaymaker, O., 1989. Holocene disequilibrium of sediment yield in British Columbia.Nature 327, 452–454.

Colls, A.E., Stokes, S., Blum, M.D., Straffin, E., 2001. Age limits on the Late Quaternary evolution of the upper Loire River. Quaternary Science Reviews 20, 743-750.

Cordier, S., Harmand, D., Losson, B., Beiner, M., 2004. Alluviation in the Meurthe and Moselle valleys (Eastern Paris Basin, France): Lithological contribution to the study of the Moselle capture and Pleistocene climatic fluctuations, Quaternaire, 15, 65-76.

Cordier, S., Harmand, D., Frechen, M., Beiner, M., 2006 - Fluvial system response to Middle and Upper Pleistocene climate change in the Meurthe and Moselle valleys (Eastern Paris basin and Rhenish Massif), Quaternary Science Reviews, 25, 1460-1474.

Cordier, S., Harmand, D., Lauer, T., Voinchet, P., Bahain, J.J., Frechen, M., 2012. Geochronological reconstruction of the Pleistocene evolution of the Sarre valley (France and Germany) using OSL and ESR dating techniques. Geomorphology 165-166, 91-106.

Cordier, S., Frechen, M., Harmand, D., 2014. Dating fluvial erosion: fluvial response to climate change in the Moselle catchment (France, Germany) since the Late Saalian. Boreas, 143, 450-468.

Cossart, E., 2008. Landform connectivity and waves of negative feedbacks during the paraglacial period, a case study: the Tabuc subcatchment since the end of the Little Ice Age, Massif des Écrins, France. Géomorphologie: relief, processus, environnement 4, 249-260.

Cossart E., 2014. Des sources sédimentaires à l'exutoire : un problème de connectivité ? Prof. Thesis, Université Blaise Pascal–Clermont 2, 241 pp.

Delmas, M., 2015. The last maximum ice extent and subsequentdeglaciation of the Pyrenees: an overview ofrecent research. Cuadernos de Investigación Geográfica 41, 109–137.

Delmas, M., Calvet, M., Gunnell, M., Braucher, R., Bourlès, D., 2011. Palaeogeography and ^{10}Be exposure-age chronology of Middle and Late Pleistocene glacier systems in the northern Pyrenees: implicationsfor reconstructing regional palaeoclimates. Palaeogeography, Palaeoclimatology, Palaeoecology 305, 109–122.

Delmas, M., Braucher, R., Gunnell, Y., Guillou, V., Calvet, M., Bourles, D., Aster Team, 2015. Constraints on Pleistocene glaciofluvial terrace age and related soil chronosequence features from vertical ^{10}Be profiles in the Ariège River catchment (Pyrenees, France). Global and Planetary Change 132, 39-53.

Demoulin, A., Mather, A., Whittaker, A., this issue. Fluvial archives, a valuable record of vertical crustal deformation.

De Winter, I.L., Storms, J.E.A., Overeem, I., 2012. Numerical modeling of glacial sediment production and transport during deglaciation. Geomorphology, 167–168, 102-114.

Dühnforth, M., Densmore, A.L., Ivy-Ochs S., Allen, P.A., 2008. Controls on sediment evacuation from glacially modified and unmodified catchments in the eastern Sierra Nevada, California.Earth Surface Processes and Landforms 33, 1602–1613.

Dvareckas V., 2000. Development of river valleys in Lithuania. Prace Geograficzne 105, 321-328.

Ehlers, J., 1996. Quaternary and Glacial Geology: Chichester, Wiley.

Fabel D., Harbor J., Dahms D., James A., Elmore D., Horn L., Daley K., Steele C. 2004. Spatial patterns of glacial erosion at a valley scale derived from terrestrial cosmogenic ^{10}Be and ^{26}Al concentrations in rock. Annals of the Association of American Geographers 94, 241–255.

Flageollet, J.-C., 1988. Quartäre Vereisungen in den lothringischen Vogesen: Anzahl, Ausdehnung unt Alter. Eiszeitalter und Gegenwart 38, 17-36.

Flageollet, J.-C., 2002. Sur les traces des glaciers vosgiens. CNRS Editions, 212 pp.

Fontana, A., Mozzi, P., Bondesan, A., 2008. Alluvial megafans in the Venetian-Friulian Plain (north-eastern Italy): evidence of sedimentary and erosive phases during the Late Pleistocene and Holocene. Quaternary International, 189, 71-90.

Fontana, A., Mozzi, P., Marchetti, M., 2014. Alluvial fans and megafans along the southern side of the Alps. Sedimentary Geology 301, 150-171.

Fuchs, M., Owen, L.A., 2008. Luminescence dating of glacial and associated sediments: review, recommendations and future directions. Boreas 37, 636-659.

García-Ruiz, J. M., Martí-Bono, C., Peña-Monné, J.L., Sancho, C., Rhodes, E., Valero, B., Gonzalez Samperiz, P., Moreno, A., 2013. Glacial and fluvial deposits in the Aragón Valley, central western Pyrenees: chronology of the Pyrenean late Pleistocene glaciers. Geografiska Annaler: Series A, Physical Geography 95, 15–32.

Gibbard, P.L., Turner, C., West, R.G., 2013. The Bytham river reconsidered. Quaternary International 292, 15-32.

Giraudi, C., Bodrato, G., Lucchi, M. R., Cipriani, N., Villa, I. M., Giaccio, B., Zuppi, G. M., 2011. Middle and Late Pleistocene glaciations in the Campo Felice Basin (central Apennines, Italy). Quaternary Research 75, 219-230.

Granger, D.E., Fabel, D., Palmer, A.N., 2001. Pliocene–Pleistocene incision of the Green River, Kentucky, determined from radioactive decay of cosmogenic ^{26}Al and ^{10}Be in Mammoth Cave sediments: Geological Society of America Bulletin 113, 825-836.

Hallet, B., Hunter, L., Bogen, J., 1996. Rates of erosion and sediment evacuation by glaciers: a review of field data and their implications. Global and Planetary Change 12, 213–235.

Hanson, P.R., Mason, J.A., Goble, R.J., 2006. Fluvial terrace formation along Wyoming's Laramie Range as a response to increased late Pleistocene flood magnitudes. Geomorphology 76, 12-25.

Harmand D., 2007. Révision du système des terrasses alluviales de la Sarre entre Sarrebourg (France, Lorraine) et Konz (Allemagne, Rhénanie-Palatinat). In : Milieux naturels et paléomilieux des Vosges au bassin de la Sarre. Revue Géographique de l'Est. Tome XLVII, 4, 235-262.

Harmand, D., Cordier, S., 2012. The Pleistocene terrace staircases of the present and past rivers downstream from the Vosges Massif (Meuse and Moselle catchments). Netherlands Journal of geosciences-GeologieenMijnbouw, 91, 91-109.

Harbor, J., Warburton, J., 1993. Relative rates of glacial and non-glacial erosion in alpine environments. Arctic and Alpine Research 25, 1–7.

Hein, A.S., Hulton, N.R.J., Dunai, T.J., Schnabel, C., Kaplan, M.R., Naylor, M., Xu, S., 2009. Middle Pleistocene glaciation in Patagonia dated by cosmogenic-nuclide measurements on outwash gravels. Earth Planetary Science Letters 286, 184–197.

Hein, A.S., Dunai, T.J., Hulton, N.R.J., Xu, S., 2011. Exposure dating outwash gravel to determine the age of the greatest Patagonian glaciations.Geology 39, 103–106.

Hétu, B., Gangloff, P., 1989. Dépôts glaciaires du Pléistocène inférieur sur le piémont des Pyrénées Atlantiques. Zeitschrift für Geomorphologie 33, 384–403.

Hétu, B., Gangloff, P., Courchesne, F., 1992. Un till de déformation du Pléistocene inférieur à la base de la formation de Lannemezan (Piémont des Pyrénées Atlantiques, France). Quaternaire 3, 53–61.

Heyman, J., Stroeven, A.P., Harbor, J.M., Caffee, M.W., 2011. Too young or too old: Evaluating cosmogenic exposure dating based on an analysis of compiled boulder exposure ages. Earth Planetary Science Letters 302, 71-80.

Hubschman, J., 1975. L'évolution des nappes alluviales antérissiennes de la Garonne dans l'avant-pays molassique. Bulletin de l'Association Française pour l'Etude du Quaternaire 12, 149–169.

Hubschman, J., 1984. Glaciaire ancien et glaciaire récent: analyse comparée de l'altération de moraines terminales nord-pyrénéennes. Montagnes et Piémonts, Hommage à François Taillefer. Revue Géographique des Pyrénées et du Sud-Ouest, Toulouse, 313–332.

Hughes, P.D., Woodward, J.C., Gibbard, P.L., Macklin, M.G., Gilmour, M.A. and Smith, G.R., 2006. The glacial history of the Pindus Mountains, Greece. The Journal of Geology 114, 413-434.

Hughes, P.D., Woodward, J.C., Van Calsteren, P.C., Thomas, L.E. and Adamson, K.R., 2010. Pleistocene ice caps on the coastal mountains of the Adriatic Sea. Quaternary Science Reviews 29, 3690-3708.

Ivy-Ochs, S., Briner, J.P., 2014. Dating disappearing ice with cosmogenic nuclides. Elements 10, 351–356.

Jackson, L.E., MacDonald, G.M., Wilson, M.C., 1982. Paraglacial origin for terraced river sediments in Bow Valley, Alberta. Canadian Journal of Earth Sciences 19, 2219-2231.

Jacobson, R.B., Elston, D.P., Heaton, J.W., 1988. Stratigraphy and magnetic polarity of the high terrace remnants in the upper Ohio and Monongahela rivers in West Virginia, Pennsylvania, and Ohio. Quaternary Research 29, 216–232.

Jordan, P., Slaymaker, O., 1991. Holocene sediment production in Lillooet River basin: a sediment budget approach. Geographie physique et Quaternaire 45, 45–57.

Kasse, C., 2014. Fluvial response to rapid high-amplitude lake-level changes during the Late Weichselian and early Holocene, Ain valley, Jura, France. Boreas 43, 403–421.

Kessler, M.A., Anderson, R.S., Stock, G.M., 2006. Modeling topographic and climatic control of east-west asymmetry in Sierra Nevada glacier length during the Last Glacial Maximum. Journal of Geophysical Research 111, F02002, doi:10.1029/2005JF000365.

Kessler, M.A., Anderson, R.S., Briner, J.P., 2008. Fjord insertion into continental margin driven by topographic steering of ice. Nature Geosciences 1, 365-369.

Koppes, M.N., Montgomery, D.R., 2009. The relative efficacy of fluvial and glacial erosion over modern to orogenic timescales. Nature Geoscience 2, 644–647.

Létolle, R. &Mainguet, M. (1993). Aral. Springerverlag, Paris, 357 p.

Lewis, C.J., McDonald, E.V., Sancho, C., Peña, J.L., Rhodes, E.J., 2009. Climatic implications of correlated Upper Pleistocene and fluvial deposits on the Cinca and Gállego Rivers (NE Spain) based on OSL dating an soil stratigraphy. Global and Planetary Change 67, 141–152.

Li Y., Harbor J., Stroeven A.P., Fabel D., Kleman J., Fink D., Caffee M., 2005. Ice sheet erosion patterns in valley systems in northern Sweden investigated using cosmogenic nuclides. Earth Surface Processes and Landforms 30, 1039–1049.

Madritsch, H., Preusser, F., Fabbri, O., 2012. Climatic and tectonic controls on the development of the River Ognon terrace system (eastern France). Geomorphology, 151–152 (2012) 126–138.

Maizels, J.K., 1979. Proglacial aggradation and changes inbraided channel patterns during period of glacier advance: analpine example. Geografiska Annaler 61, 1-2, 87-101.

Mandier, P., 1984. Le relief de la moyenne vallée du Rhône au Tertiaire et au Quaternaire, essai de synthèse paléogéographique. Document du BRGM n° 151, 871 pp.

MARGO Project Members, 2009. Constraints on the magnitude and patterns of ocean cooling at the Last Glacial Maximum. Nature Geosciences 2, 127–132.

Marks, L., 2004. Middle and Late Pleistocene fluvial systems in central Poland. Proceedings of the Geologists Association 115, 1–8.

Marquette, G.C., Gray, J.T., Gosse, J.C., Courchesne, F., Stockli, L., Macpherson, G., Finkel, R., 2004. Felsenmeer persistence under non-erosive ice in the Torngat and Kaumajet mountains, Quebec and Labrador, as determined by soil weathering and cosmogenic nuclide exposure dating. Canadian Journal of Earth Sciences 41, 19–38.

Marren, P.M., Toomath, S.C., 2013. Fluvial adjustements in response to glacier retreat: Skaftafelljökull, Iceland. Boreas 42, 57-70.

Matoshko A., 2004. Evolution of the fluvial system of the Pypriat, Desna and Dnieper during the Late Middle- Late Pleistocene. Quaternaire 15, 117-128.

Niviere, B., Lacan, P., Regard, V, Delmas, M., Calvet, M., Huyghe, D., Roddaz, B., 2016. Evolution of the late Pleistocene Aspe River (Western Pyrenees, France).Signature of climatic events and active tectonics. C.R. Géosciences, 203-212.

Occhietti, S., 1990. Lithostratigraphie du Quaternaire de la vallée du Saint-Laurent : méthodes, cadre conceptuel et séquences sédimentaires. Géographie physique et Quaternaire 44, 137-145.

Occhietti, S., Richard, P.J.H., 2003. Effet réservoir sur les âges 14C de la Mer de Champlain à la transition Pléistocène-Holocène : révision de la chronologie de la déglaciation au Québec méridional. Géographie physique et quaternaire 57, 115-138.

Occhietti, S., Cordier, S., Harmand, D, Kulinicz, E., 2012. Lithostratigraphie des terrasses de la Meurthe et de la Sarre en périphérie des Vosges et de l'Hunsrück (France, Allemagne) : réponse des cours d'eau aux fluctuations climatiques pléistocènes. Géomorphologie : relief, processus, environnement 4, 441-458.

Oetelaar, G.A., 2002. River of change: a model for the development of terraces along theBow River, Alberta. Géographie Physique et Quaternaire 56, 155–169.

Olley, J., Caitcheon, G., Murray, A.S., 1998. The distribution of apparent dose as determined by Optically Stimulated Luminescence in small aliquots of fluvial quartz: Implications for dating young sediments. Quaternary Science Reviews 17, 1033-1040.

Owczarek, P., Nawrot, A., Migala, K., Malik, I., Korabiewski, B., 2014. Flood-plain responses to contemporary climate change in small High-Arctic basins (Svalbard, Norway). Boreas 43, 384-402.

Owen, L.A., Sharma, M.C., 1998. Rates and magnitudes of paraglacial fan formation in the Garhwal Himalaya: implications for landscape evolution. Geomorphology 26, 171–184.

Panin, A., Adamiec, G., Filippov, V., 2015. Fluvial response to proglacial effects and climate in the upper Dnieper valley (Western Russia) during the Late Weichselian and the Holocene. Quaternaire 26, 27-48.

Panin, A., Adamiec, G., Buylaert, J.P., Matlakhova, E., Moska, P., Novenko, E., this issue. Two Late Pleistocene climate-driven incision/aggradation rhythms in the middle Dnieper River basin, west-central Russian Plain.

Panzer,W, 1926. Talentwicklung und Eiszeitklima in nord-östlichen Spanien. Abh. Der Senckenbergischen Naturforschenden Gesellschaft, 39, 141-182.

Parent, M., 1987. Late Pleistocene stratigraphy and events in the Asbestos-Valcourt Region, Southeastern Quebec. Digitized Theses. Paper 1664, 356 pp.

Parent, M., Occhietti, S. (1988). Late Wisconsinian and Champlain sea invasion in the St-Lawrence valley, Québec.Géographie physique et quaternaire, Vol. 42, n° 3, 215 – 246.

Passmore, D.G., Waddington, C., 2009. Paraglacial adjustment of the fluvial system to Late Pleistocene deglaciation: The Milfield Basin, northern England. In Knight, J. and Harrison, S., (eds). Periglacial and paraglacial processes and environments. Geological Society, London, Special Publications 320, 145-164.

Pazzaglia, F.J., 2013. Fluvial Terraces. *In* Schroder J.F., Wohl, E. Treatise of Geomorphology. New York, Elsevier, 379-412.

Peña, J.L., Sancho, C., Lewis, C., McDonald, E., Rhodes, E., 2004. Datoscronologicos de las morrenas terminales delglaciardel Gallego y su relacion con las terrazas fluvioglaciares (Pirineo de Huesca). GeografiaFisica de Aragon, Aspectos generales y tematicos 71–84.

Peña Monné J.L., Turu V., Calvet M. 2011.Les terrasses fluvials del Segre i afluents principals : descripció d'afloraments i assaig de correlació. In : Turu V. and Constante A. (eds), El Cuaternario en España y áreas afines, avances en 2011, XIII Reunión Nacional de Cuaternario, Andorra, 4-7 juillet, Asociación Española para el Estudio del Cuaternario (AEQUA), 51-55.

Penck, A. (Translated L. Braemer), 1885. La période glaciaire dans les Pyrénées. Bulletin de la Société d'Histoire Naturelle de Toulouse 19, 105–200.

Penck, A., Brückner, E., 1909. Die Alpen im Eiszeitalter.Tauchnitz, 1199 pp.

Preusser, F., Graf, H.R., Keller, O., Krayss, E., Schlüchter, C., 2011. Quaternary glaciation history of northern Switzerland. Eiszeitalter und Gegenwart 60, 282-305.

Putkonen, J., Swanson, T., 2003. Accuracy of cosmogenic ages for moraines. Quaternary Research 59, 255–261.

Putkonen, J., O'Neal, M.A., 2006. Degradation of unconsolidated Quaternary landforms in the western North America.Geomorphology 75, 408–419.

Reille, M., Andrieu, V., 1993. Variations de la limite supérieure des forêts dans les Pyrénées (France) pendant le Tardiglaciaire. Compte rendu de l'Académie des Sciences de Paris Série II 316, 547–551.

Ritter, D.F., Ten Brink, N.W., 1986. Alluvial fan development and the glacial-glaciofluvial cycle. Nenana Valley, Alaska. Journal of Geology 94, 613-615.

Rixhon, G., Briant, B., Cordier, S., Duval, M., Jones, A., Scholz, D., this issue. Revealing the pace of river landscape evolution during the Quaternary: recent developments in numerical dating methods.

Roussel, E., Chenet, M., Grancher, D., Jomelli, V., 2008. Processus et rythmes de l'incision des sandar proximaux postérieure au petit âge glaciaire (sud de l'Islande). Géomorphologie : relief, processus, environnement 4, 235-248.

Sancho, C., Peña, J.L., Lewis, C., McDonald, E., Rhodes, E., 2003. Preliminary dating of glacial and fluvial deposits in the Cinca River Valley (NE Spain): chronological evidences for the Glacial Maximum in the Pyrenees? In: Ruiz Zapata, M.B., Dorado Valiño, M., Valdeolmillos

Rodríguez, A., Gil García, M.J., BardajíAzcárate, T., De Bustamante Gutiérrez, I., Martínez Mendizábal, l, (Eds.), Quaternary Climatic Changes and Environmental Crises in the Mediterranean Region. Universidad de Alcalá de Henares, Ministerio de Ciencia y Tecnología, 169–173.

Sancho, C., Peña, J.L., Lewis, C., McDonald, E., Rhodes, E., 2004. Registros fluviatiles y glaciares cuaternarios en las cuencas de los ríos Cinca y Gállego (Pirineos y depresióndel Ebro). In: Colombo Piñol, F., Liesa Carrera, C.L., MeléndezHevia, G., Pocoví Juan, A., Sancho Marcén, C., Soria de Miguel, A.R. (Eds.), Geo-Guías 11 Itinerarios Geológicos por Aragón, 181–216.

Sancho, C., Calle, M., Peña-Monne, J.L., Duval, M., Oliva-Urcia, B., Pueyo, E.L., Benito, G., Moreno, A., 2017. Dating the Earliest Pleistocene alluvial terrace of the Alcanadre River (Ebro Basin, NE Spain): Insights into the landscape evolution and involved processes. Quaternary International, doi.org/10.1016/j.quaint.2015.10.050

Seret, G., 1966. Les systèmes glaciaires du bassin de la Moselle et leurs enseignements. Revue royale belge de géographie, 2, 3, 577 pp.

Seret, G., Dricot, E., Wansard, G., 1990. Evidence for an early glacial maximum in the French Vosges during the last glacial cycle.Nature 346, 453-456.

Shackleton, N. J. (1987). Oxygen Isotopes, Ice volume and Sea Level. Quaternary Science Reviews, 6, p 183-190.

Schildgen, T., Dethier, D.P., Bierman, P., Caffee, M., 2002. 26Al and 10Be dating of late Pleistocene and Holocene fill terraces: a record of fluvial deposition and incision, Colorado Front Range. Earth Surface Processes and Landforms 27, 773– 787.

Smedley, R.K., Glasser, N.F., Duller G.A.T., 2016. Luminescence dating of glacial advances at Lago Buenos Aires (~46°S), Patagonia. Quaternary Science Reviews 134, 59-73.

Sörgel, W., 1939. Das Diluviale System. Fortschritte Geologie und Paleontologie 12, 1-92.

Stange, K.M., van Balen, R., Carcaillet, J., Vandenberghe, J., 2013. Terrace staircase development in the Southern Pyrenees Foreland: inferences from ^{10}Be terrace exposure ages at the Segre River. Global and Planetary Change 101, 97–112.

Stange, K.M., van Balen, R.T., Kasse, C., Vandenberghe, J., Carcaillet, J., 2014. Linking morphology across the glaciofluvial interface: A ^{10}Be supported chronology of glacier advances and terrace formation in the Garonne River, northern Pyrenees, France. Geomorphology 207, 71–95.

Starkel, L., 1994. Reflection of the glacial–interglacial cycle in the evolution of the Vistula river basin, Poland. Terra Nova, 6, 486–494.

Starkel, L., 2003. Climatically controlled terraces in uplifting mountain areas. Quaternary Science Reviews 22, 2189–2198.

Starkel, L., Gębica, P., Superson, J., 2007. Last Glacial–Interglacial cycle in the evolution of river valleys in southern and central Poland. Quaternary Science Reviews 26, 2924-2936.

Starkel, L., Michczynska D., Gebica P., Kiss T., Panin, A., Persoiu, I., 2015. Climatic fluctuations reflected in the evolution of fluvial systems of Central-Eastern Europe (60-8 ka cal BP). Quaternary international 388, 97-118.

Straffin, E.C., Blum, M.D., Colls, A., Stokes, S. (1999). Alluvial stratigraphy of the Loire and Arroux Rivers (Burgundy, France). Quaternaire 10, 271-282.

Trasher, I.M., Mauz, Chiverrell, R.C., Lang, A., 2009. Luminescence dating of glaciofluvial deposits : A review. Earth-Science Reviews 97, 133-146.

Turu, V., Calvet, M., Bordonau, J., Gunnell, Y., Delmas, M., Vilaplana, J.M., Jalut, G., 2016. Did Pyrenean glaciers dance to the beat of global climatic events? Evidence from the Würmian sequence stratigraphy of an ice-dammed palaeolake depocentre in Andorra. In : Hughes P.D. and Woodward J.C.(eds), Quaternary Glaciation in the Mediterranean Mountains, Geological Society, London, Special Publication 433-6.

Van Balen, R. T., Busschers, F. S., Tucker, G. E., 2010. Modeling the response of the Rhine-Meuse fluvial system to Late Pleistocene climate change. Geomorphology 114, 440–452.

Van den Berg, M. W., 1996. Fluvial sequences of the Maas, a 10 Ma record of neotectonics and climate change at various timescales. Thèse, Université de Wageningen, 181 p.

Vandenberghe, J., 1995. Timescales, climate and river development. Quaternary Science Reviews, 14, 631–638.

Vandenberghe, J., 2001. A typology of Pleistocene cold-based rivers. Quaternary International 79, 111-121.

Vandenberghe J., 2003. Climate forcing of fluvial system development: an evolution of ideas. Quaternary Science Reviews 22, 2053–2060.

Vandenberghe J., 2008. The fluvial cycle at cold-warm-cold transitions in lowland regions: a refinement of theory. Geomorphology 98, 275–284.

Vandenberghe, J., 2012. Multi-proxy analysis: a reflection on essence and potential pitfalls. Netherlands Journal of Geosciences - Geologie en Mijnbouw 91, 263-269.

Vandenberghe, J., 2014. River terraces as a response to climatic forcing: Formation processes, sedimentary characteristics and sites for human occupation. Quaternary International 370, 3-11.

Wallinga, J., 2002. Optically stimulated luminescence dating of fluvial deposits: a review. Boreas, 31, 303-322.

Wiederkehr, E., Dufour, S., Piégay, H., 2010. Localisation et caractérisation semi-automatique des géomorphosites fluviaux potentiels. Exemples d'applications à partir d'outils géomatiques dans le bassin de la Drôme (France) », Géomorphologie : relief, processus, environnement 2, 175-188.

White, T.S., Bridgland, D.R., Howard, A.J., White, M.J., 2010. Evidence of the Trent terrace archive for lowland glaciation of Britain during the Middle and Late Pleistocene. Proceedings of the Geologists' Association 121, 141–153.

White, T.S., Bridgland, D.R., Westaway, R., Straw, A., 2016. Evidence for late Pleistocene glaciation of the British margin of the Southern North Sea. Journal of Quatern ary Science DOI: 10.1002/jqs.2826.

White, T.S., Bridgland, D.R., Limondin-Lozouët, N., Schreve, D.C., this issue. Fossils from Quaternary fluvial archives : sources of biostratigraphical, biogeographical and palaeoclimatic evidence.

Whiteman, CA., Rose, J., 1992. Thames river sediments of the British Early and Middle Pleistocene. Quaternary Science Reviews 11, 363-375.

Wilkie, K., Clargue, J.J., 2009. Fluvial response to Holocene glacier fluctuations in the NostetukoRiver valley, southern Coast Mountains, British Columbia.In : Knight, J. & Harrison, S. (Eds) Periglacial and Paraglacial Processes and Environments. Geological Society, London, Special Publications 320, 199–218.

Woodward, J.C., Hamlin, R.G.B., Macklin, M.G., Hughes, P.D., Lewin, J., 2008. Glacial activity and catchment dynamics in northwest Greece: Long-term river behaviour and the slackwater sediment record for the last glacial to interglacial transition. Geomorphology 101, 44-67.

Žebre, M., Stepišnik, U., 2015. Glaciokarst landforms and processes of the southern Dinaric Alps. Earth Surface Processes and Landforms 40, 1493-1505.

Zreda, M.G., Phillips, F.M., Elmore, D., 1994. Cosmogenic ^{36}Cl accumulation in unstable landforms, 2, Simulations and observations on eroding moraines. Water Resources Research 30, 3127–3136.

Figures

Figure 1: European glaciated regions discussed in the text. Ice margins relate to maximum known ice extents, and do not always correspond to the last glacial maximum (LGM). See text for details.

Figure 2: Datings on Pyrenean fluvial terraces (modified from Calvet, 2004). 1—a: Last glaciation (Würmian) maximumice extent (MIE); b:Middle Pleistocene ice extent. 2—Dated terrace staircase and/or glacio-fluvial complex.3 — Authors and dating method.

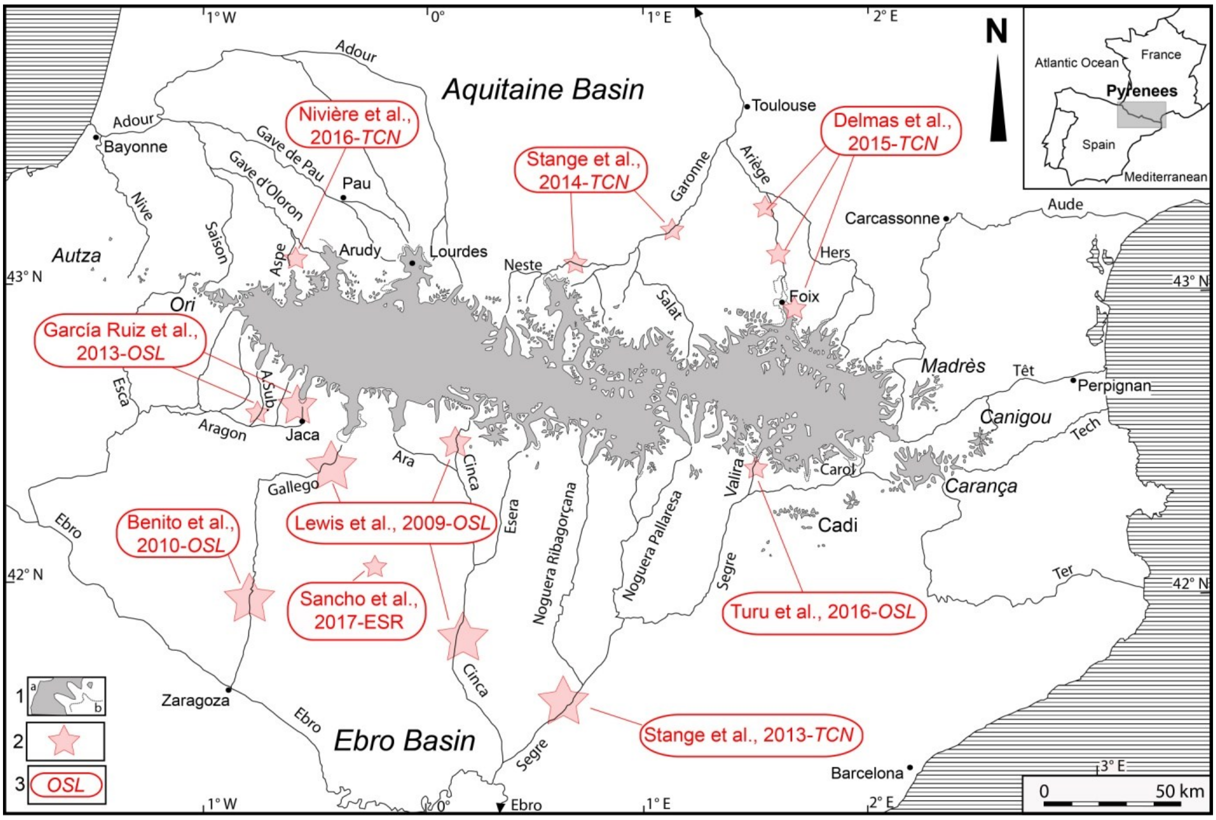

Figure 3: The Vosges Massif and surrounding area, a key place for the study of the glacial-fluvial coupling

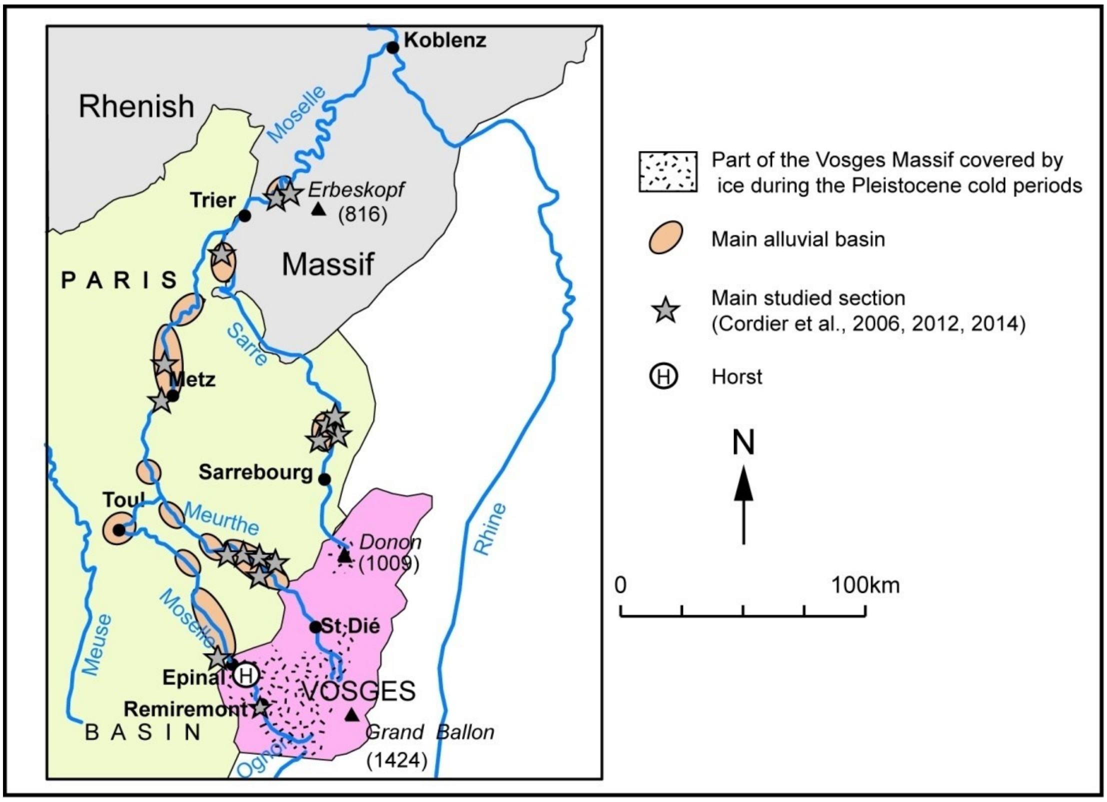

Figure 4 : The sections of A-Vathiménil (Meurthe terrace Me4, +30 m relative height) and B-Golbey-Pré Droué (Moselle terrace M3, + 20 m) show a clear erosive contact between the lower and upper units. The lower unit is allocated to a glacial period on the basis of sedimentology (Vathiménil) and OSL dating (Golbey-Pré Droué, MIS 6 age). The erosive contact between both units is allocated to the melting of the Vosges glacier (‘proglacial erosion’). The upper unit (allocated to the MIS 5 age at Golbey on the basis of OSL dating) contains a significant proportion of sediments from the glaciated areas : their deposition likely corresponds to the paraglacial reworking of the sediments from the upper Moselle and Meurthe catchments.

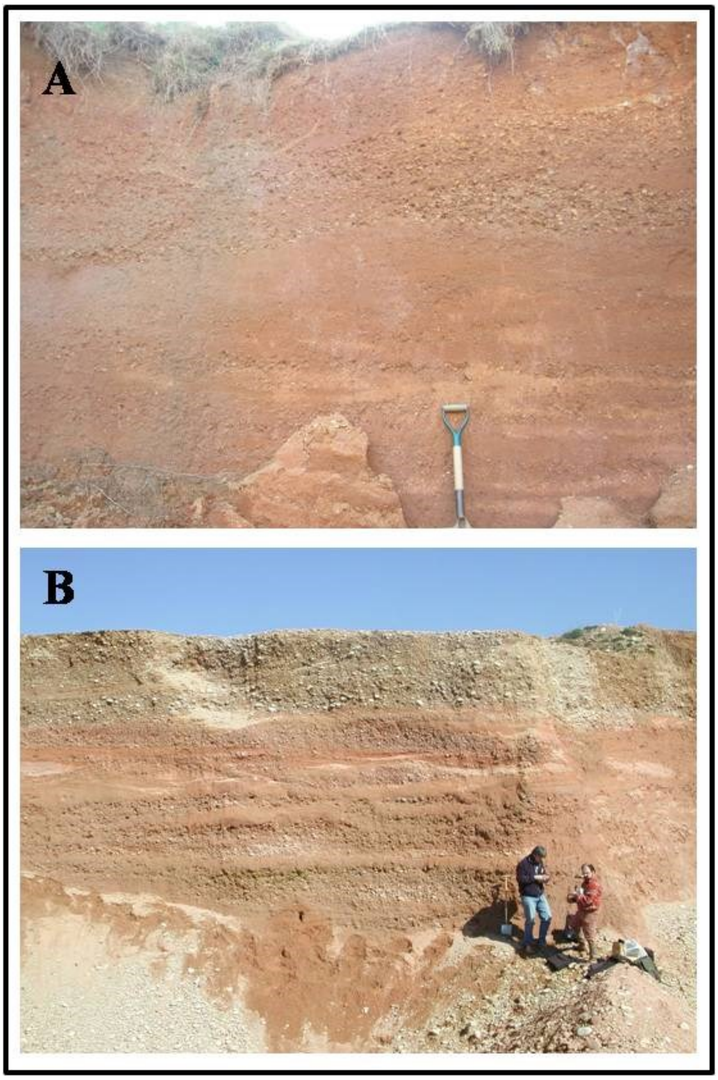

Figure 5: Schematic diagram of the fluvial response to glacial dynamics during the deglaciation (see text for discussion).

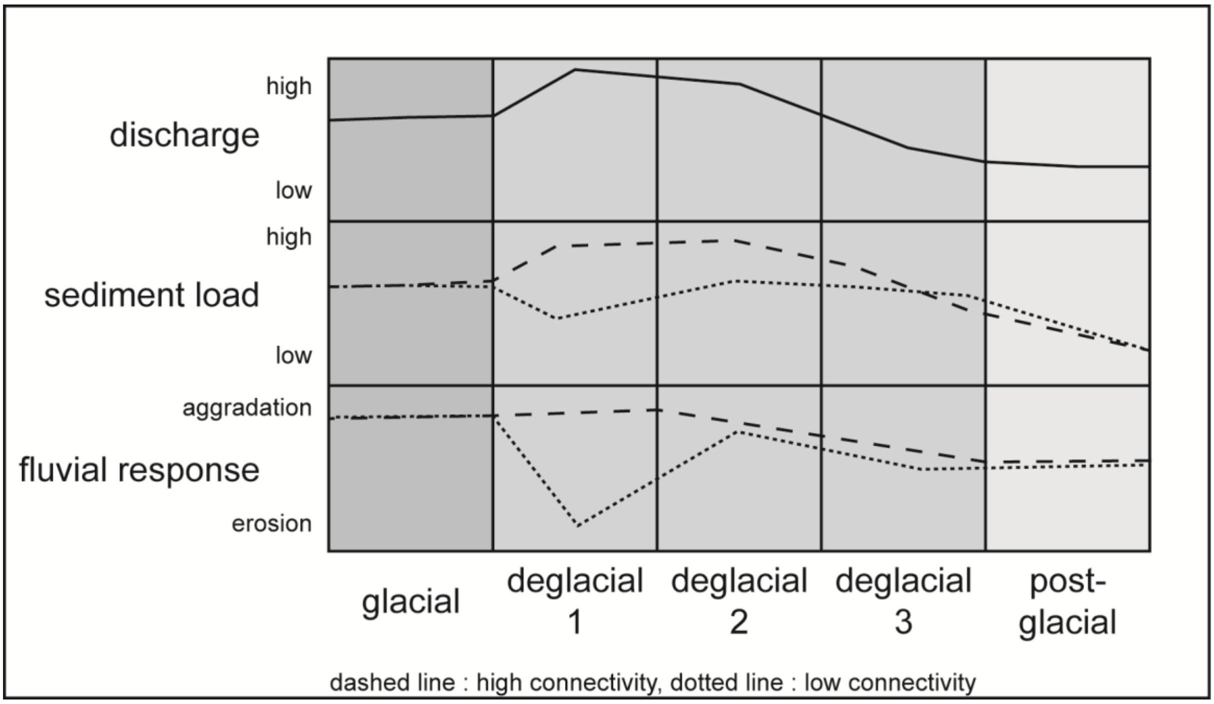

Table 1: Correlations between pyrenean fluvio-glacial terraces. M in white squares mean undated or indirectly dated moraines, M in grey squares mean dated moraines. T in white squares mean undated or indirectly dated moraines, T in grey squares mean dated moraines.

	NORTHERN PYRENEES			SOUTHERN PYRENEES				
valley	**Aspe/Ossau**	**Garonne**	**Ariège**	**Valira/Sègre**	**Cinca/ Alcanadre**	**U. Gallego**	**L. Gallego**	**U. Aragon**
Upper Pleistocene	Aspe T1 «Gurmençon» +40m 18±2 ka TCN Ossau T1 «Ogeu» $\boxed{M}$	T1 «Basse plaine» $\boxed{M}$ +25-14m Rivière, $14.6^{+9.6}_{-4.3}$ ka Cazères, $13^{+6.7}_{-3.9}$ ka (TCN, from upstream to downstream)	T1 «Grausse de Pamiers» $\boxed{M}$ +14-25-xxm Montgaillard, $17.5^{+2.0}_{-3.5}$ ka Filaliter, $13.8^{+3.6}_{-0.4}$ ka Cintegabelle, $13^{+3.5}_{-0.5}$ ka (TCN, from upstream to downstream)	TQ7, +3-5m (down) TQ6, +8-14m (down) $\boxed{M}$ SVT8/T9, +8-10m (ups) 32.8±1.2 ka OSL TQ5,+16-26m (down) $\boxed{M}$ SVT7, +15m (ups) TQ4,+35-47m (down) $61.8^{+4.6}_{-5.4}$ ka TCN TQ3,+48-65m 99.6^{+31}_{-19} ka TCN	Qt9, +6-10m 11 ka OSL 15-22 ka ^{14}C Qt8, +20m $\boxed{M}$ 47-51 ka OSL Qt7, +35-50m $\boxed{T}$ 61 ka OSL Qt6, +60m 97 ka OSL	«Lower terrace» +35-45m $\boxed{M}$ 32-45 ka OSL «Middle terrace» +11-17m $\boxed{T}$ 66-103 ka OSL	T12, +3-10m 16.8±1.3 ka OSL T11, 10-12m 55.4±7.4 ka OSL 54.4±8.8 ka OSL T10, +20m 124±13 ka OSL 110±20 ka OSL	$\boxed{M}$ 20m terrace 68±7 ka OSL
Middle Pleistocene	Aspe T2 «Agnos» $\boxed{T}$ Ossau T2 «Herrères» T3/T4	T2 «basse terrasse» «Blagnac-Seysse» $\boxed{T}$ T3 «moyenne terrasse» Léguevin-St Lys T4 «haute terrasse» Rieumes	T2 «Basse Boulbonne» Tournac, 60-145 ka (minimum age) $\boxed{T}$ T3 «Haute Boulbonne» Ch.Fiche, 204-226 ka (minimum age) T4	SV-T5, +40m (ups) $\boxed{T}$ 125±11 ka OSL 120±15 ka OSL TQ2,+77-88m (down) $138.8^{+46.7}_{-22.8}$ ka TCN TQ1,+100-113m (down) $202^{+35.3}_{-32.9}$ ka TCN SVT3/SVT4, +105m (downstream) SV-T2, +80m $\boxed{T}$ (upstream) ?	Qt5, +80m 178 ka OSL Qt4 Qt3 ---- B/M (750ka) ---- Qt2	«Upper terrace» +46-72m $\boxed{T}$ 151 ka OSL ------------	T09, +30-40m 147±16 ka OSL 133±10 ka OSL 163±22 ka OSL 181±13 ka OSL 156±26 ka OSL T08, +45m T07 ---- B/M (750ka) ---- T06, +60m T05, +75m T04, +85m T03, +95m	$\boxed{M}$ 60m terrace 263±4.8 ka OSL ----------
Lower Pleistocene -Pliocene	T5/Lanemezan megafan top $\boxed{T}$	T5 «Très haute terrasse» «Hte Bouconne, cailloutis de Lomagne» Lanemezan megafan	Lannemezan-high gravel on plateaus	TQ0, +125-140m SVT1, +140-170m (upstream) ?	Alcanadre Qt1, +160m 1276±104 ka ESR Reverse paleomag.		T02, +105m T01	
Ref.	Nivière et al., 2016 Hubschman, 1984	Stange et al., 2014 Hubschmann, 1975	Delmas et al., 2015 Hubschmann, 1975	Stange et al., 2013 (TQ) Turu et al. 2016, Pena et al., 2011 (SVT)	Lewis et al., 2009 Sancho et al., 2017	Lewis et al., 2009	Benito et al., 2010	Garcia Ruiz et al., 2010