Spatial and temporal variability in sedimentation rates associated with cut-off channel infill deposits: Ain River, France

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[1] Floodplain development is associated with lateral accretion along stable channel geometry. Along shifting rivers, the floodplain sedimentation is more complex because of changes in channel position but also cutoff channel presence, which exhibit specific overflow patterns. In this contribution, the spatial and temporal variability of sedimentation rates in cutoff channel infill deposits is related to channel changes of a shifting gravel bed river (Ain River, France). The sedimentation rates estimated from dendrogeomorphic analysis are compared between and within 14 cutoff channel infills. Detailed analyses along a single channel infill are performed to assess changes in the sedimentation rates through time by analyzing activity profiles of the fallout radionuclides ^{137}Cs and unsupported ^{210}Pb. Sedimentation rates are also compared within the channel infills with rates in other plots located in the adjacent floodplain. Sedimentation rates range between 0.65 and 2.4 cm a^{-1} over a period of 10 to 40 years. The data provide additional information on the role of distance from the bank, overbank flow frequency, and channel geometry in controlling the sedimentation rate. Channel infills, lower than adjacent floodplains, exhibit higher sedimentation rates and convey overbank sediment farther away within the floodplain. Additionally, channel degradation, aggradation, and bank erosion, which reduce or increase the distance between the main channel and the cutoff channel aquatic zone, affect local overbank flow magnitude and frequency and therefore sedimentation rates, thereby creating a complex mosaic of sedimentation zones within the floodplain and along the cutoff channel infills. Last, the dendrogeomorphic and ^{137}Cs approaches are cross validated for estimating the sedimentation rate within a channel infill.

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1. Introduction

[2] A wide range of sedimentation rates have been reported for numerous floodplains, both in Europe [*Walling and He*, 1997; *Hensel et al.*, 1999; *Jeffries et al.*, 2003] and in the United States [*Friedman et al.*, 1996; *Kleiss*, 1996; *Hupp*, 2000; *Ross et al.*, 2004]. These sedimentation rates are mainly estimated using (1) dendrochronology when trees are available for coring [*Hupp*, 2000], (2) artificial markers [*Kleiss*, 1996; *Noe and Hupp*, 2005], (3) radionuclide profiles [*Walling and He*, 1997; *Aalto et al.*, 2003], but also at a shorter timescale by (4) repetitive topographical surveys [*Hooke*, 1995].

[3] Floodplain sedimentation rates vary in space and time [*Schumm and Lichty*, 1965]. They decrease from the channel bank to the floodplain as a function of roughness, and associated decreasing suspended sediment concentration. These high sedimentation rates are generally found on floodplain areas bordering the river channel, reflecting the importance of diffusion of coarser particles from the channel onto the floodplain during floods [*Walling and He*, 1998]. This spatial structure has been widely exemplified, first, by identifying the specific environment of deposition in lowland rivers, distinguishing natural levees from backswamps [*Hupp*, 2000] and second, by integrating the lateral distance as an independent variable in statistical models [*Walling and He*, 1998] or by representing physical gradient in suspended sediment concentration in overbank flood waters [*Pizzuto*, 1987].

[4] Sedimentation rate can also be related to the age of the floodplain surface, because older areas, which have undergone a longer period of sedimentation are higher in elevation and therefore less frequently inundated with sediment-laden water [*Nanson and Beach*, 1977; *Hooke*, 1995]. Overbank flow depth is a key parameter in floodplain sedimentation models [*Nicholas and Walling*, 1997; *Walling and He*, 1998] which is dynamically associated with the magnitude and frequency of overbank flows [*Moody and Troutman*, 2000]. *Aalto et al.* [2003] reported along the Beni and Mamore rivers discrete packages of sediment of uniform age, 20–80 cm thick, with 8-year

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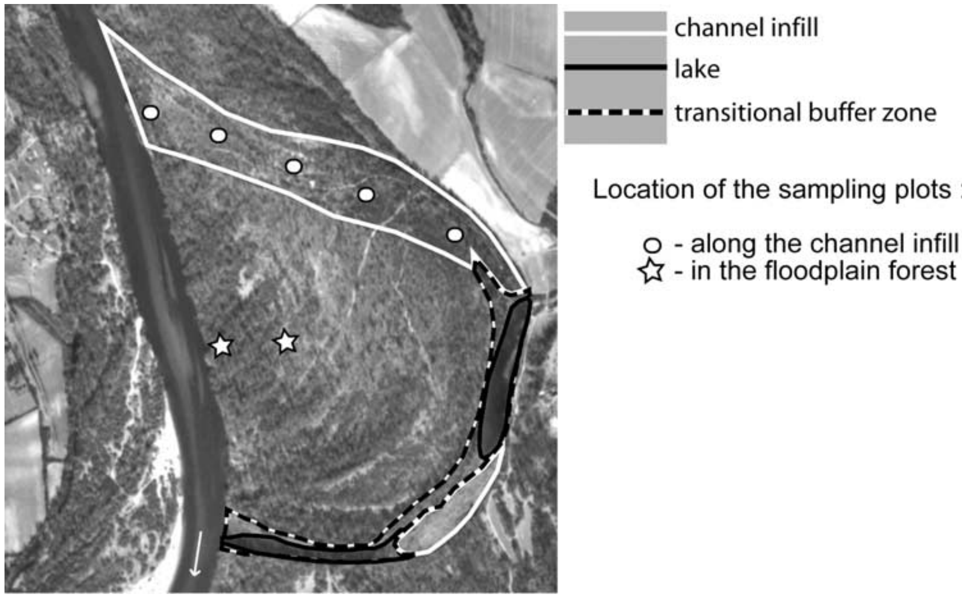

Figure 1. Aerial photograph of a characteristic cutoff channel, the Planet on the Ain, showing the three subunits associated with longitudinal water conditions with the terrestrial channel infill upstream and the lake downstream in contact with each other by transition semiaquatic areas.

recurrence intervals of rapidly rising floods showing how floodplain deposition is also related to the rapidity of flood stage rise.

[5] Such well-known spatial and temporal trends can be modified by inner floodplain heterogeneity associated with variation in microtopography and vegetation density and flexibility [*Leopold et al. 1964*; *Jacobson and Coleman*, 1986]. Spatial variation in vegetation cover influences the roughness and flow velocities on floodplains [*Arcement and Schneider*, 1989] and affects depositional conditions [*Nanson and Beach*, 1977; *Piégay 1997*]. However, the vegetation effect is local and does not seem to counteract the key factors at the reach scale (e.g., distance to the bank; magnitude and frequency of flood [*Steiger et al.*, 2001a]). Woody debris distribution in the channel also controls the magnitude and frequency of flooding at a local scale and the conveyance of flow along specific floodplain paths that may explain spatial variation in sedimentation rate on floodplains [*Jeffries et al.*, 2003].

[6] Floodplain features are also not homogeneous in space in relation to associated processes of sedimentation. Floodplain formation is accomplished by both vertical and lateral accretion [*Leopold et al.*, 1964; *Nanson and Croke*, 1992; *Hupp*, 2000]. Vertical accretion is the dominant process that creates floodplain surfaces [*Ross et al.*, 2004], and occurs during overbank flows when reduction of critical water velocity allows the deposition of sediment [*James*, 1985]. Sediment is carried by diffusion due to a gradient in turbulence between the channel and the inundated floodplain [*Pizzuto*, 1987] but also by advective transport allowing sand distribution onto the floodplain. Lateral accretion becomes important during channel migration and point bar extension, which creates complex deposits with sequences of ridges and swales within the developing floodplain as the result of divergent flows in the meander bends [*Nanson and Beach*, 1977; *McKenney et al.*, 1995]. Vertical accretion is critical along lowland floodplains on which spatial and temporal trends in sedimentation have been widely demonstrated, whereas lateral accretion associated with bed load sand is more typical of piedmont and intramontane actively shifting rivers [*Nanson and Beach*, 1977; *Hupp*, 2000]. Nevertheless, other local processes may also be influential in vertical accretion [*Nanson and Croke*, 1992]. Abandoned channel accretion which can represent a significant portion of actively migrating river floodplains (e.g., 20% [*Lewis and Lewin*, 1983]), is another process of floodplain formation with a sedimentation gradient from fine lacustrine material in the upper part of the deposit to relatively coarse/sandy material near the base [*Nanson and Croke*, 1992].

[7] Despite the aforementioned understanding of floodplain dynamics, few reports document sedimentation rates in cutoff channels [*Shields and Abt*, 1989; *Piégay et al.*, 2002]. A cutoff channel is a channel that has been abandoned because of channel migration. They are termed oxbows (along meandering rivers) or dead arms, backwaters or sloughs (along relatively high-gradient rivers). Cutoff channels usually have three main units (Figure 1): (1) the central aquatic area, also termed the cutoff lake [*Timms*, 1992], which supports characteristic aquatic vegetation and is principally responsible for high biodiversity within the cutoff channel; (2) the channel infill deposit that isolates the lake from the main channel during at least low flow stages, at the upstream end and sometimes at the downstream end [*Erskine et al.*, 1982]; and (3) a transitional semiaquatic buffer zone, fringing the lake, of varying size that may occur with greatest development at either end of the lake, though typically upstream. The channel infill deposit is supplied by bed load at the early stages of the cutoff channel, and once vegetated, by relatively fine-grained overbank sediments similar to other parts of the floodplain [*Allen*, 1965; *Erskine and Melville*, 1983].

[8] The geometry of the former channel, particularly the angle θ between the old bendway mouth and cut channel

mouth controls the channel infill deposition [*Shields and Abt*, 1989]. Most authors suggest that the thickness of the alluvial fill in the former channel is dependent upon age and proximity to the main channel; older former channels in close proximity to the main channel may develop the greatest amount of fill [*Bradley and Brown*, 1992; *Erskine et al.*, 1992; *Hooke*, 1995]. Similarly, sedimentation rates tend to be greatest during the early stages of cutoff filling; the magnitude and frequency of flooding in the cutoff channel progressively decreases as the landform fills [*Hooke*, 1995].

[9] In a context of shifting channels, the processes of channel infill sedimentation should be more complex because of the nonstationary hydraulic conditions at the channel, floodplain boundaries and of advective processes controlling sand deposits along the channel infill pathways. As a consequence, the objectives of this paper are to evaluate the spatial and temporal variability of sedimentation rates within and among channel infills of a shifting gravel bed river. We focus on upstream channel infills over former channels located in the Ain River valley, France (Figure 2) and overbank sedimentation once the channel infills are developed and forested. In this study, sediment load, hydrological regime and flood frequency are considered as constant regardless of the sampling area because the studied reach has no major tributaries modifying the upstream conditions and we analyzed the variability in sedimentation rate in relation to local geomorphic conditions.

[10] This work which specifically focused on sedimentation rates constitutes an additional piece of results to the previous contributions provided by *Piégay et al.* [2000] who highlighted the effects of channel shifting on sedimentation patterns in cutoff channel lakes and its consequences on the regeneration on successional patterns of aquatic vegetation community or by *Piégay et al.* [2002] who worked on cutoff channel infill morphology showing the complex structure of habitat conditions (surficial grain size, overbank sedimentation structure, vegetation community, soil characters) in relation with former channel geometry (braided versus meandering pattern).

2. Characteristics of the Study Reach

[11] The Ain River, a tributary to the Rhône River, drains a catchment of 3672 km^2, located mainly within the low-elevation limestone Jura Mountains, northwest of the French Alps. This course can be divided into two main segments: the upper segment is about 160 river km long, extending from the headwaters to Pont d'Ain (Figure 2) and flows in a V-shaped valley, and the lower segment, which is 40 river km long, and extends from Pont d'Ain to the confluence with the Rhône, flows across a 8 km wide alluvial plain (see Figure 2 on which the Holocene corridor is drawn).

[12] The river channel in this lower segment has a sinuous pattern with free-meandering subreaches (e.g., Bublane-Villette and Blyes reaches, Figure 2), surrounded by many former channels which have mostly been cut off during the past 200 years [*Citterio*, 1997]. This lower segment has a relatively steep slope (0.12% to 0.18%) and is characterized by high, short-lived flood peaks, which largely occur during the winter. The mean annual discharge of the river at the gauging station of Chazey-sur-Ain (record from 1960 to 2006) is 123 m^3 s^{-1}, with the 10-year flood having a discharge of about 1740 m^3 s^{-1}.

[13] The lower segment is considered to be among the most picturesque, unpolluted, and uncontrolled river reaches in France [*Marston et al.*, 1995]. Riparian areas along the lower Ain support several related but contrasting diverse habitats (particularly in the aquatic zone), whose existence and longevity are associated with channel mobility and cutoff dynamics [*Pautou and Girel*, 1986]. The present research has direct bearing on the long-term conservation of these ecologically important riparian systems.

[14] At least 150 lakes occur within the 0.5 to 2 km-wide lower-valley floodplain [*Bornette and Amoros*, 1990]. They vary in size, age, distance to main channel, and elevation relative to the main channel. Two reaches contain most of the lakes; one 7 km reach located 12 km above the confluence with the Rhône and another $\sim$15 km reach between Mollon and Villette-sur-Ain (Figure 2). The physical characteristics of these lakes vary seasonally and from one site to another depending on the original planform of the cutoff channel and its hydrological connectivity with both the main channel and the alluvial aquifer. From 1954 to 2000, the shift in channel pattern toward a single channel in relation with changes in bed load supply and floodplain resistance due to both catchment and floodplain spontaneous reforestation facilitated the generation of the most recent cutoff channels (Figure 2d) [*Liébault and Piégay*, 2002].

[15] Three principal river patterns (braided, wandering, and meandering forms) succeeded each other along the Ain floodplain during the 20th century. We use the conventional definitions of meandering and braided channels [*Leopold and Wolman*, 1957; *Church*, 1992]. Following *Kellerhals and Church* [1989], wandering pattern is used to describe an intermediate pattern between braided and true meandering patterns, where the channel is too narrow to maintain a braided pattern yet side and midchannel bars may occur with a dominant and a secondary channel. Each of these patterns develop particular cutoff features: large and sinuous abandoned channels in the case of the meandering pattern, straight long, narrow, and steep abandoned channels in the case of the braided pattern; straight, wide, and relatively short in the wandering pattern.

[16] The river underwent a complex vertical evolution over the last 3 decades. In the upper part, the channel has degraded 1 m because of the propagation of a sediment deficit from the dam at Allement (12 km upstream from Pont d'Ain) built in 1960. In other sections downstream not yet affected by this sediment deficit (e.g., sections upstream of meander bends), the changes in channel pattern from braided to meandering provide local sediment inputs with aggradation [*Rollet et al.*, 2005].

3. Materials and Methods

[17] Among lakes mapped by *Bornette and Amoros* [1990], we selected 14 of the largest, representing a range of age and geometry. We identified each of their upstream terrestrial channel infills using 7 series of air photos available between 1945 and 2000. Four sites had an original morphology of meandering channels, two of braided channels, and eight of wandering channels. For this last group, the lakes are not across the whole channel but across

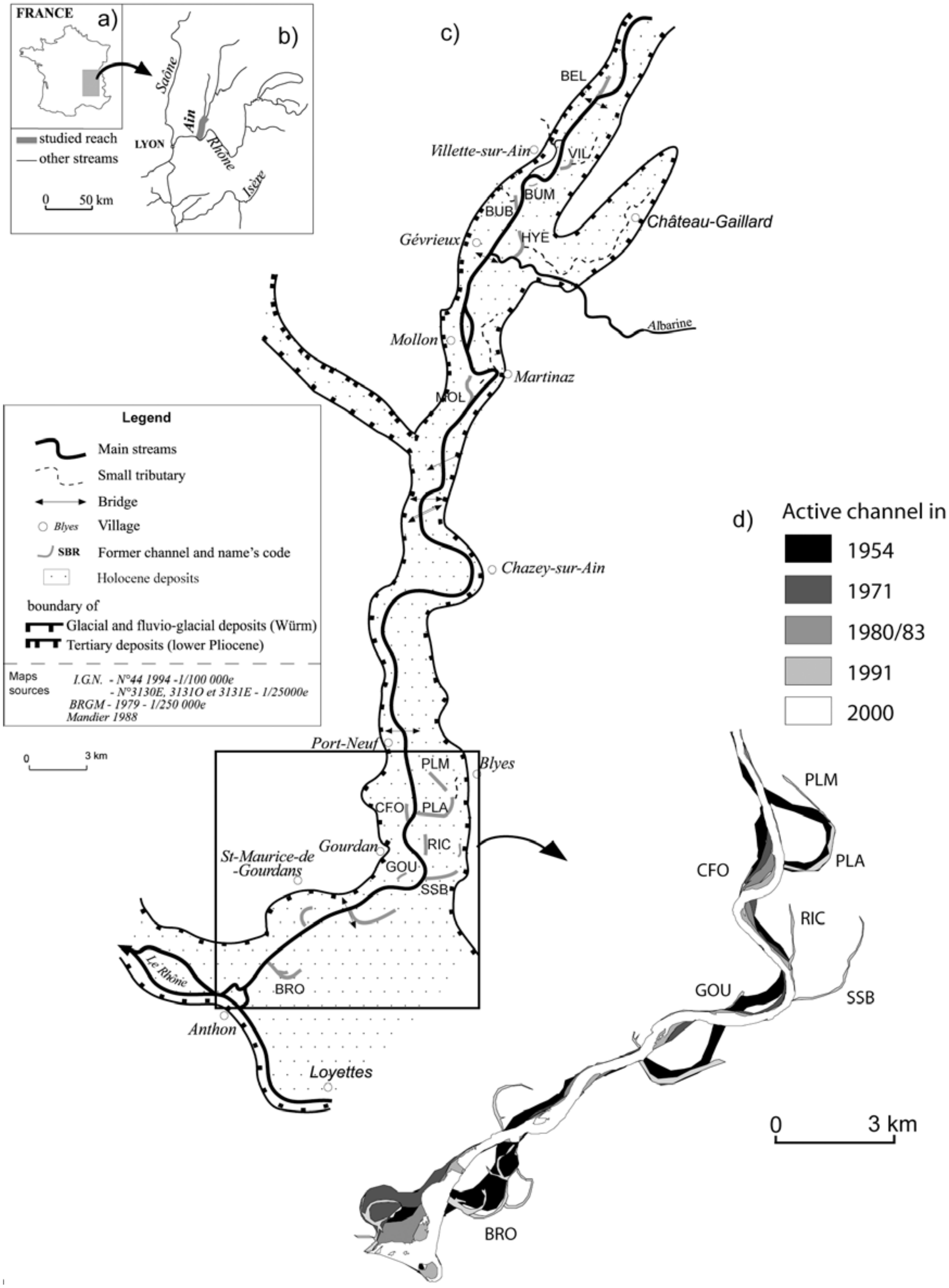

Figure 2. Location of the Ain valley (a) in France and (b) in the Rhône catchment, and location of (c) the channel infills studied in the Ain Valley. (d) Illustration of the channel mobility between 1954 and 2000 in the reach Blyes-Anthon.

backwater features, and isolated from the main channel by the downstream end of a bar. The cutoff channels range in age from 23 to 90 years, with the date of cutoff determined from air photo analysis, old map observations, and field discussion with landowners. The sedimentation rates were compared between and within these channel infills. Detailed analyses along a single channel infill, at the Ricotti site (Figure 2, labeled "RIC") were performed to assess changes in sedimentation rates through time. Additionally, we compared sedimentation rates on channel infills with sedimentation rates in forest plots located on the neighboring floodplain, using tree cores from 66 and 104 trees, respectively.

3.1. Dendrochronological Analysis of Sediment Deposition Rate

[18] The field routine, at 14 channel infill sites and 6 floodplain sites, consisted of the establishment of two to six plots where detailed dendrogeomorphic analyses (tree ring dating) were conducted. These plots were located such that they represented along-channel infill conditions and sedimentation trends from near the river bank to the lake. For the comparison between the 6 floodplain sites and the 4 associated channel infill sites, we used a specific sampling design, with a pair of plots (one located 10 to 15 m from the bank; one located ca 100 m from the bank).

[19] Dendrogeomorphic techniques provide the net or minimum rate of channel infill sediment (and, in a few cases, erosion). It is most of the time the "gross" deposition but we are never sure that there has not been some scour or erosion of soil above the roots between the time of germination and when we measured the depth of burial so that we always consider it as a minimum.

[20] Typically six or more trees were sampled at each plot (10 to 61 trees per channel infill site); a total of 377 trees were sampled for the channel infill survey. Replication at each plot is necessary to account for local depositional variation and to ensure the determination of a mean rate with an acceptable standard error (SE < mean) [*Hupp*, 1988; *Hupp and Bornette*, 2003]. The procedure follows that of *Sigafoos* [1964] where the specimen trees are partly excavated down to the top of the normally horizontally radiating root mass, a level that is established at the time of germination. The depth of burial from the top of major roots to the present ground surface provides a conservative estimate of net sediment deposition during the life of the tree. The tree is then cored near its base with an increment borer; the extracted core is returned to the laboratory for cross dating and age determination. The depth of burial above the root collar is divided by the age of the tree to provide an estimate of net deposition rate. Depth to roots was measured at a distance away from the trunk that is approximately the trunk diameter at ground surface to avoid the collar curvature. This technique has been shown to be internally consistent and has been used with considerable success along streams in the Southeastern USA [*Hupp and Morris*,1990; *Hupp et al.*, 1993; *Ross et al.*, 2004; *Jolley and Lockaby*, 2006] and the Great Plains [*Friedman et al.*, 1996; *Scott et al.*, 1997] regions of the United States. *Fraxinus excelsior* was preferentially sampled because it was the most common species at the sampling plots, they have easily determined ring boundaries, do not experience multiple and/or missing rings to the degree of other species, and they have a well developed collar curvature with flat roots near the trunk. *Populus nigra* and *Alnus glutinosa* were also sampled occasionally when *Fraxinus* was not common at a plot.

3.2. Assessment of Controlling Factors of Sedimentation Rates

[21] We conducted an analysis of different series of air photos which have been scanned, mosaicked, georeferenced and corrected in planimetry to measure lateral channel mobility near the upstream ends of the channel infills and to determine the surface area, length, and age of each channel infill [*Piégay et al.*, 1999]. Each plot was located precisely on the GIS layer and their distance from the upstream end (river bank) of the channel infill determined. For dating cutoffs older than 1945, we used archival maps and written reports [*Rollet et al.*, 2005]. A similar approach was taken for the adjacent floodplain plots.

[22] We also surveyed a long profile of the Ain River in April 1999 along the low-flow channel and compared it with a previous survey in 1976. These data were used to determine the present and past channel gradient and vertical channel stability (quantitative amounts of channel aggradation or degradation) near the upstream end of the channel infills.

[23] We designated 145 elementary (250 m long) channel segments from the GIS for which we extracted average values of lateral and vertical channel mobility among the existing surveys. At the upstream end of each channel infill, a crest stage gauge was installed to determine the channel stage above which the channel infill flows in reference to the hydrological record provided by the gauging station at Chazey-sur-Ain (1960 to present). A similar approach was taken for the adjacent floodplain plots.

3.3. Assessment of Temporal Trend in Sedimentation Rates

[24] We determined the mean depth of overbank fine sediments above the bed load deposits of the channel infill at each plot with a hand auger. The infill is characterized by bed load deposits during the early stage of development and later by overbank fine sediment once vegetation is established [*Allen*, 1965; *Erskine and Melville*, 1983; *Erskine et al.*, 1992]. Here, we studied only the second fine-grained stage after the trees were established. We divided this depth by the age of the channel infill to get a long-term deposition rate and compared it with the relatively short-term rates provided by dendrogeomorphic analyses.

[25] Along the Ricotti channel infill (Figure 2), 58 trees were sampled for dendrogeomorphic analysis, approximately double the amount sampled at other channel infills, to provide detailed estimates of temporal trends by comparing sedimentation rates among differently aged tree cohorts. Moreover, six cores of overbank fine sediment were collected from the channel infill at sites extending from the channel bank to the lake. Each was partitioned into 2 cm sections providing 142 samples. Seventy one sections (every other 2 cm interval) were used to establish the depth distribution of ^{137}Cs and unsupported ^{210}Pb activities within these fine sediment deposits (sieved fraction <2 mm). The activities of the two radionuclides were measured simultaneously by gamma ray spectrometry using a low-background n-type HPGe detector. Count times were sufficient to provide measurements with a precision of ±ca 10% at the 95% level of confidence.

[26] Three reference soil cores were also collected in the valley bottom, from sites that have never flooded over the last 60 years (B, MG, ML, located in the Holocene plain outside the Q_{100} flood envelope). Because core B is the closest to the floodplain sampling site, the associated ^{137}Cs and ^{210}Pb inventories are assumed to represent the direct atmospheric fallout inputs of these two radionuclides to the local area. The degree to which the measured inventory exceeds the reference inventory provides a first-order esti-

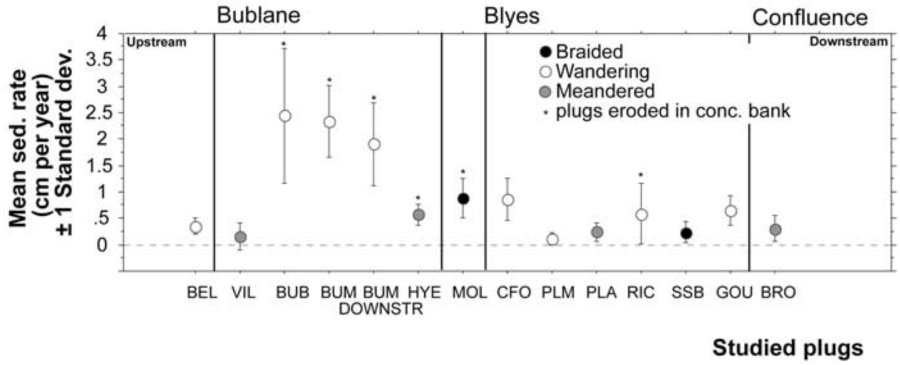

Figure 3. Distribution of the mean sedimentation rates at the studied channel infills, related to their fluvial geomorphological pattern. The channel infills are arranged from upstream to downstream (left to right). The vertical black lines indicate the major reaches. See Figure 2 for site location.

mate of the relative amount of sedimentation at the various sites.

[27] The ^{137}Cs measurements have been also used to estimate the sedimentation rates based on the 1963 peak [*Walling and He*, 1997]. In each of the ^{137}Cs activity profiles, the maximum activity has been identified as representing the 1963 peak, in order to establish the position of the floodplain surface in 1963 and the depth of sediment deposited during the period 1963–1999. This is the standard approach used for lake and floodplain sediments and we assume it is equally applicable to these infill deposits, even if the sedimentation is complex and not constant through time. A peak marking the 1963 depth is expected for two reasons. First, the sediment surface at that time would receive fallout and fallout was greatest in 1963. Second, if sediment eroded from the catchment was deposited at the site, this could also be expected to be characterized by maximum ^{137}Cs concentrations around 1963, because of the maximum fallout at this time. In the period shortly after fallout, the fresh fallout will remain at or close to the soil surface and any soil eroded from surface sources will be characterized by maximum ^{137}Cs levels.

[28] No estimates of sedimentation rate were derived from the excess ^{210}Pb core data. With the exception of cores 6, 8, and 11, the profiles do not demonstrate the expected gradual reduction in activity with depth. This suggests that it is not valid to assume a constant sedimentation rate and that variation in grain size may complicate the pattern [*He and Walling*, 1996].

[29] Additional data were also collected in 1997: 11 topographical cross sections (CS) of the channel infill located every 55 m from the bank to the lake of the former channel. The area from CS 9 to CS 11, near the upstream end of the lake largely supports herbaceous macrophytes whereas the other CS locations mainly support woody floodplain vegetation. On each cross section, we conducted dendrogeomorphic analyses and made 5 to 6 soil borings to estimate the depth of overbank silt deposits above the gravel deposits.

3.4. Data Analyses

[30] We averaged individual sedimentation rates by plot and by channel infill in order to assess interchannel infill variability and longitudinal trends within each channel infill. Box plots illustrating sedimentation rate variability at all plots for all sites were constructed. We log-transformed mean sedimentation rate per plot or channel infill to perform simple and multiple linear regressions and to predict changes in sedimentation rate in space and time. Kruskall-Wallis and Mann-Whitney tests were also used for validating statistical significance of mean differences between sets of plots and channel infills. Because all the explanatory variables were not available for every channel infill (e.g., the age of the cutoff or the local overbank discharge), statistical analyses were performed on partial samples.

4. Results

4.1. Comparisons of Sedimentation Rates Between Channel Infills

[31] The estimates of sedimentation rate obtained from dendrogeomorphic analysis averaged 0.65 cm a^{-1} for all of the Ain River channel infills, with a maximum of 2.4 cm a^{-1} for the left bank Bublane channel infill (BUM) and a minimum of 0.11 cm a^{-1} for the Old Planet channel infill (PLM) (Figure 3). The sedimentation rate is dependent on the reach in which the channel infill is located. As shown on Figure 3, the channel infills located in the Bublane reach had higher sedimentation rates than those located in the Blyes reach. In these two reaches where 5 to 6 channel infills are observed, sedimentation rates also vary markedly between each of the channel infills. We observed the highest sedimentation rates where wandering channel infill geometry occurred in conjunction with a location downstream of eroding banks (cutbanks on outside bends; Figure 3). Intermediate channel infill sedimentation rates were observed at sites where only one of these two factors (wandering geometry or cutbank erosion) occurred. The lowest sedimentation rates were observed where there was neither cutbank erosion nor wandering geometry. These observations are validated by a Kruskal-Wallis test (Figure 4). Similarly, the interplot variability in sedimentation rate within each channel infill is also the highest in the wandering channel infills located downstream of cutbanks. There are no differences in sedimentation rate where the channel infill is associated with either a meandering or a braided former channel.

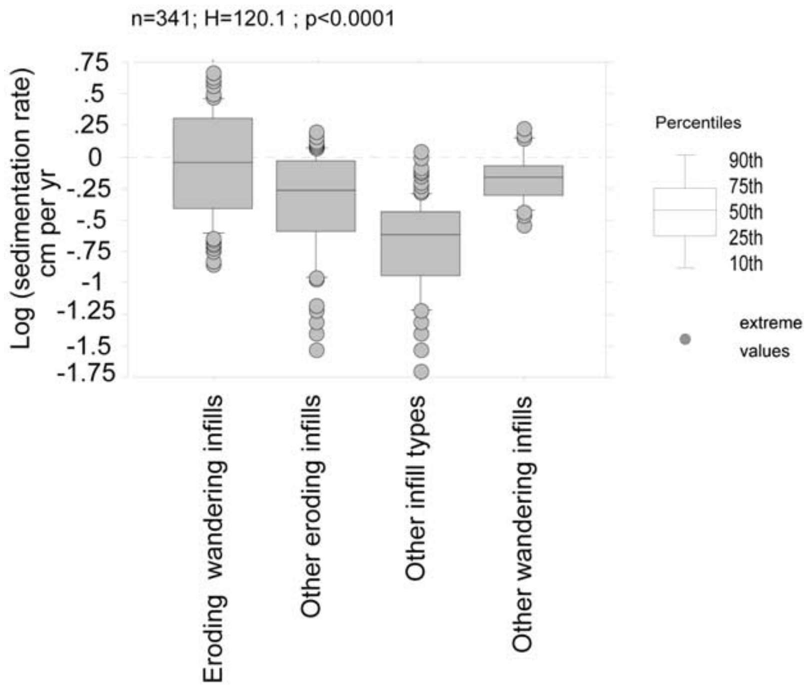

Figure 4. Distribution of the log-transformed mean sedimentation rates according to geomorphological types and bank erosion. The Kruskall Wallis test indicates that the differences among the groups are statistically significant at the probability of 0.0001.

4.2. Factors Controlling Variation in Sedimentation Rates Between Channel Infills

[32] Log-transformed mean sedimentation rate per channel infill, local overbank discharge, and mean age of trees were tested for linear relations (Figure 5). An inverse linear relationship exists between sedimentation rate and the mean age of trees, ranging from 10 to 40 years (n = 13; Figure 5a), regardless of the type of former channels.

[33] An inverse linear relation also exists between sedimentation rate and local overbank discharge and several internal groupings can be identified: the wandering channel infills and the meandering and braided channel infills (n = 10; Figure 5b). For a given local overbank discharge, the wandering channel infills experience higher sedimentation rates than other channel infill types and also have younger trees than the meander or braided infills (Figure 5c). Where similar hydrologic connections exist, wandering infills exhibit higher sedimentation rates than other channel types and support younger trees. They also have a lower junction angle θ than the two other types, and also no obvious downstream gradient of sedimentation rates. The slope of the longitudinal linear trend of sedimentation rate from the bank to the lake is close to 0 or negative, indicating that the sedimentation rates at plots near the lake are slightly higher than those close to the bank. There are no differences between the three types in terms of historic lateral or vertical movement of the main channel (Figure 6). For the 6 wandering type infills, the sedimentation rate is related to tree age and channel degradation between 1976 and 1999 (Figures 5d and 5e); the older channel infills are relatively disconnected because of main channel degradation.

4.3. Comparison of Sedimentation Rates Over the Last 5 Decades (From the Dendrogeomorphic Survey) With Those for the Entire Period of Channel Infill Solution

[34] Statistical analyses reveal that there is no correlation between the log-transformed mean annual sedimentation rate per channel infill provided by the dendrogeomorphic survey over a period of 10 to 50 years with channel infill age, ranging from 40 to 90 years old (Figure 7a). Short-term sedimentation rates from the dendrogeomorphic survey are usually lower than the mean annual sedimentation rates documented for the entire (long-term) period of channel infill generation, as the quotient of the total depth of deposition above the gravel and the age of the cutoff (Figure 7b). This is consistent with the previous work showing a decrease in sedimentation rate through time but the decrease is not constant in time from one site to another so that no relation can be determined between the recent sedimentation rate and the sedimentation rate since cutoff. Observing channel infills plotted just below the 1:1 correlation there are some outliers: two with a high sedimentation rate in the recent period (BUM, BUB) and three with very low recent sedimentation rates compared to the long-term rate (PLM, VIL, SSB) (Figure 7b). VIL and SSB are rarely flooded and they have been isolated from the main channel by newly deposited floodplain, after 1956 for SSB following the cutoff of Planet (Figure 2) and after 1980 for VIL. At PLM, the dendrochronological survey was conducted relatively close to the lake because tree and shrub clearing to maintain a flow path near the main channel. In these three cases, the mean distance of plots to the channel bank is distinctly higher than for the others (test of Mann-Whitney: $z = -1.18$; $p = 0.26$).

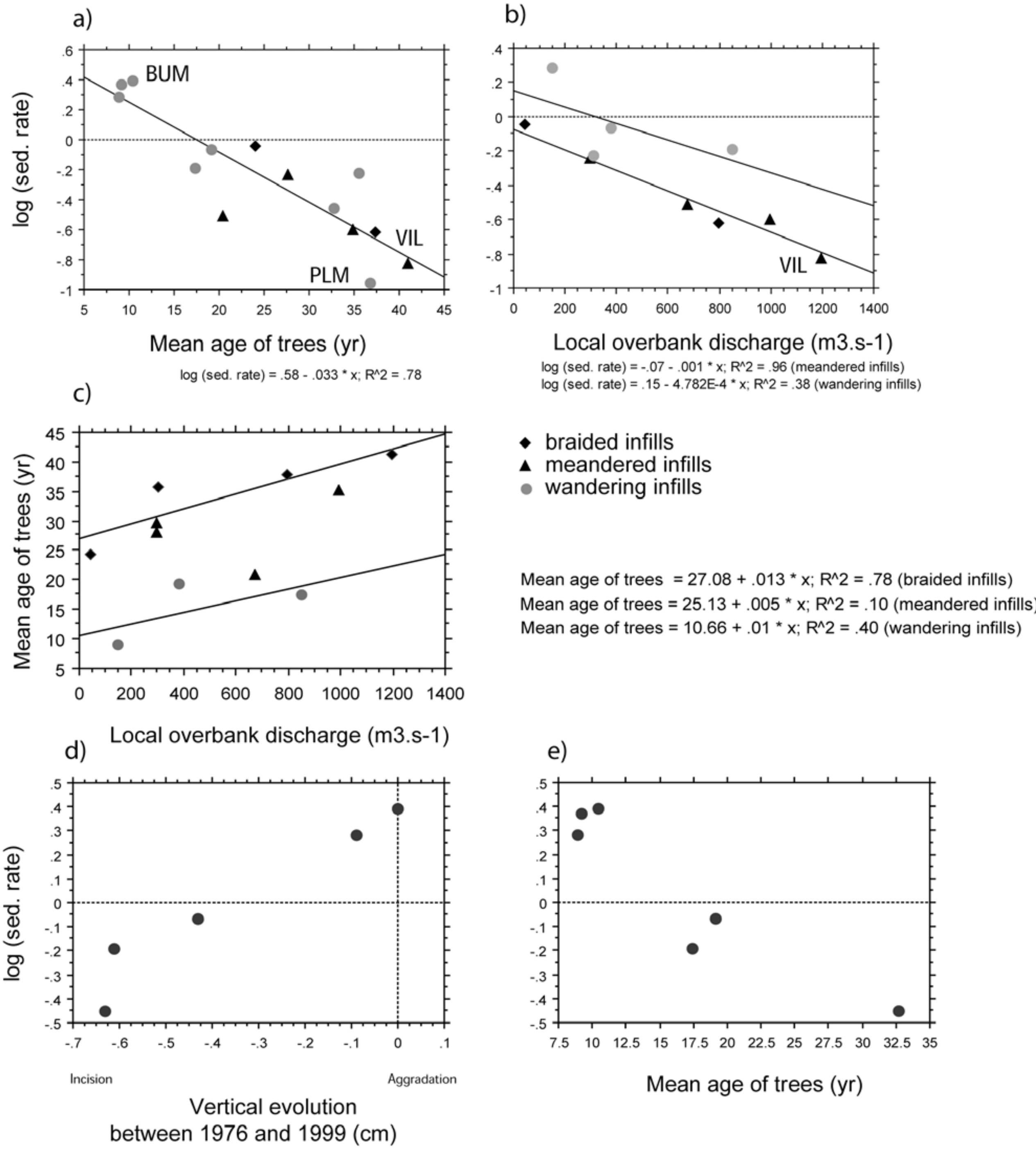

Figure 5. Scatterplot showing the link between the log-transformed mean sedimentation rate and (a) the mean age of trees cored and (b) the overbank discharge, between the mean age of trees and (c) the overbank for the three infill types, and between log-transformed mean sedimentation rate and (d) the annual degradation rate of the channel between 1976 and 1999 and (e) the mean age of trees cored for the wandering infills.

4.4. Comparisons of Sedimentation Rates Between Channel Infills and Floodplain Forest Plots

[35] The sedimentation rates on channel infills are significantly higher (mean = 1.03 cm a^{-1}) than on the adjoining floodplain plots (mean = 0.74 cm a^{-1}), because of the higher frequency of overbank flooding (Figures 8a and 8b) regardless of distance from channel bank and age of trees (Figures 8c and 8d). If the sedimentation rates are normalized by the frequency of overbank flooding, the difference is not statistically valid. The channel infills therefore have a greater hydraulic connectivity than the floodplain plots. They are flooded on average by a $\sim$500 $m^3\ s^{-1}$ flood corresponding to $Q_{0.5}$, whereas the floodplain forest plots are flooded in average by a 1250 $m^3\ s^{-1}$ flood (Q_5). The

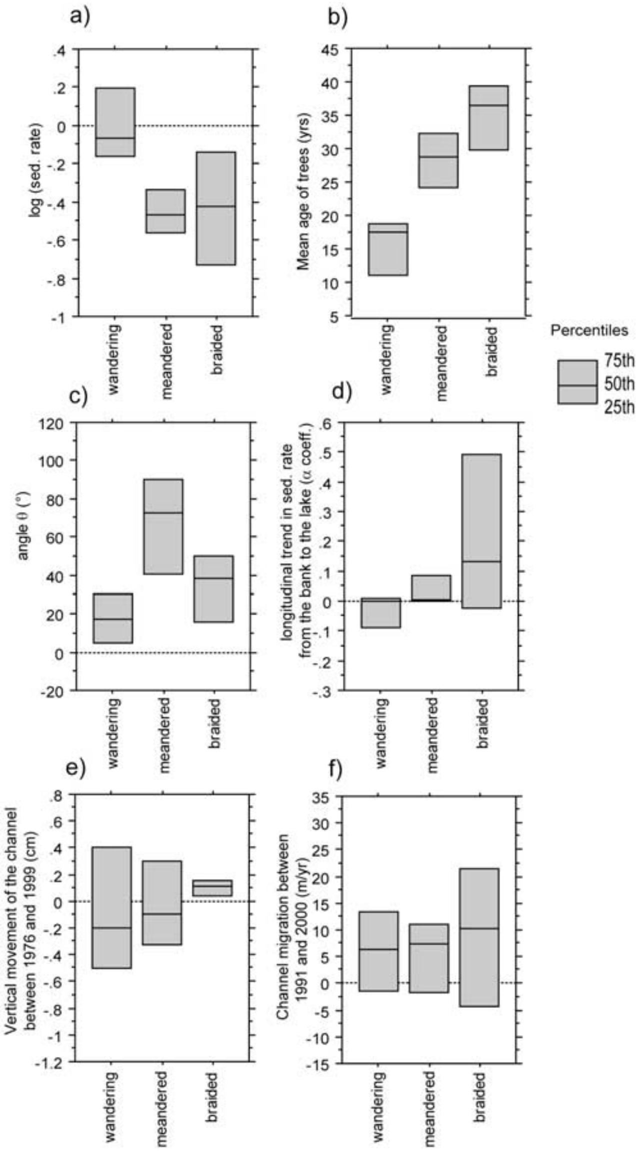

Figure 6. Comparison of the three infill types according to (a) the sedimentation rate, (b) the mean age of trees, (c) the angle θ, (d) the coefficient α of the longitudinal trend in sedimentation rate, (e) the vertical movement of the channel between 1976 and 1999, and (f) the channel migration between 1991 and 2000.

sedimentation rates observed on both the plots near the banks (e.g., 15–25 m) and those farther inside the floodplain (e.g., 100–200 m) exhibit different sedimentation patterns (Figure 8e). For floodplain plots, those located near the bank have higher sedimentation rates than those located farther inside the floodplain. This differs from the situation found for the channel infills. The channel infill plots located 100–200 m from the bank do not have substantially different sedimentation rates than the plots located near/within banks and in all cases they also have a higher sedimentation rate than the near-bank floodplain plots.

4.5. Intrachannel Infill Variability in Sedimentation Rate

[36] We investigated the downstream variation in mean sedimentation rate within each channel infill. A linear decrease in sedimentation in relation to the mean distance to the bank is observed (Figure 9a). It appears that this relationship between sedimentation rate and distance to the bank differs between meandering channel infills and wandering channel infills. A similar linear trend of decreasing sedimentation rate from the bank to the lake was observed, regardless of the type of channel infill, when analyzing the detailed information (the distance per plot and not the mean distance per channel infill). A slightly higher sedimentation rate occurs in wandering and braided channel infills with increasing distance from the bank, compared to meandering channel infills (Figure 9b). For the wandering type, observation of the residuals from the linear relation shows a link with the mean age of the trees; high residuals are positively associated with decreasing tree age. There is no relation between the residuals and mean tree age for meandering or braided channel infills (Figure 9c).

4.6. Detailed Analysis of Intrachannel Infill Variability in Sedimentation Rate for the Ricotti Channel Infill

[37] The Ricotti site (RIC) is a wandering channel infill (Figure 2d), colonized by woody vegetation between 1830

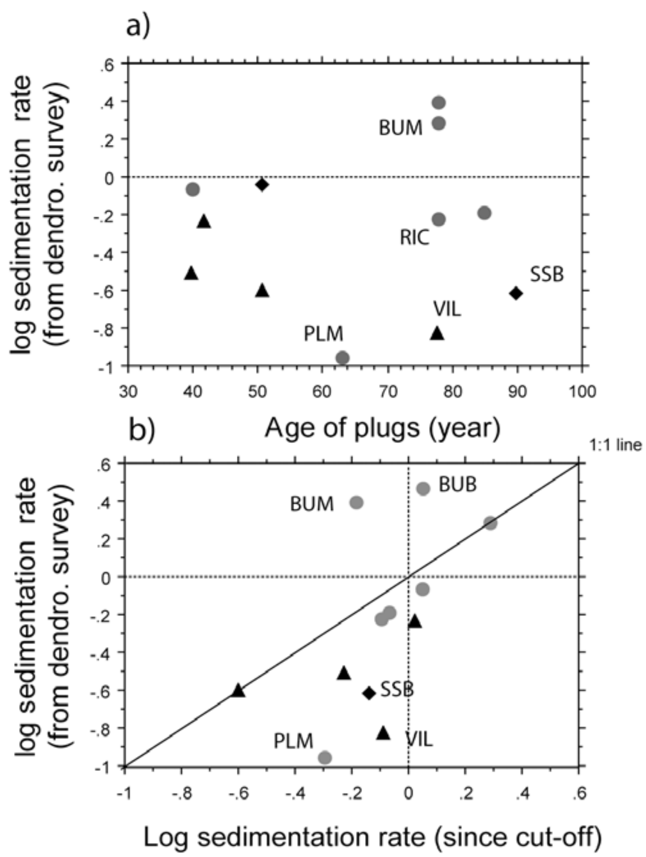

Figure 7. Scatterplot showing the link between the log-transformed mean sedimentation rate provided by the dendrogeomorphic survey with (a) the age of the former channel and (b) the log-transformed sedimentation rate since the cutoff.

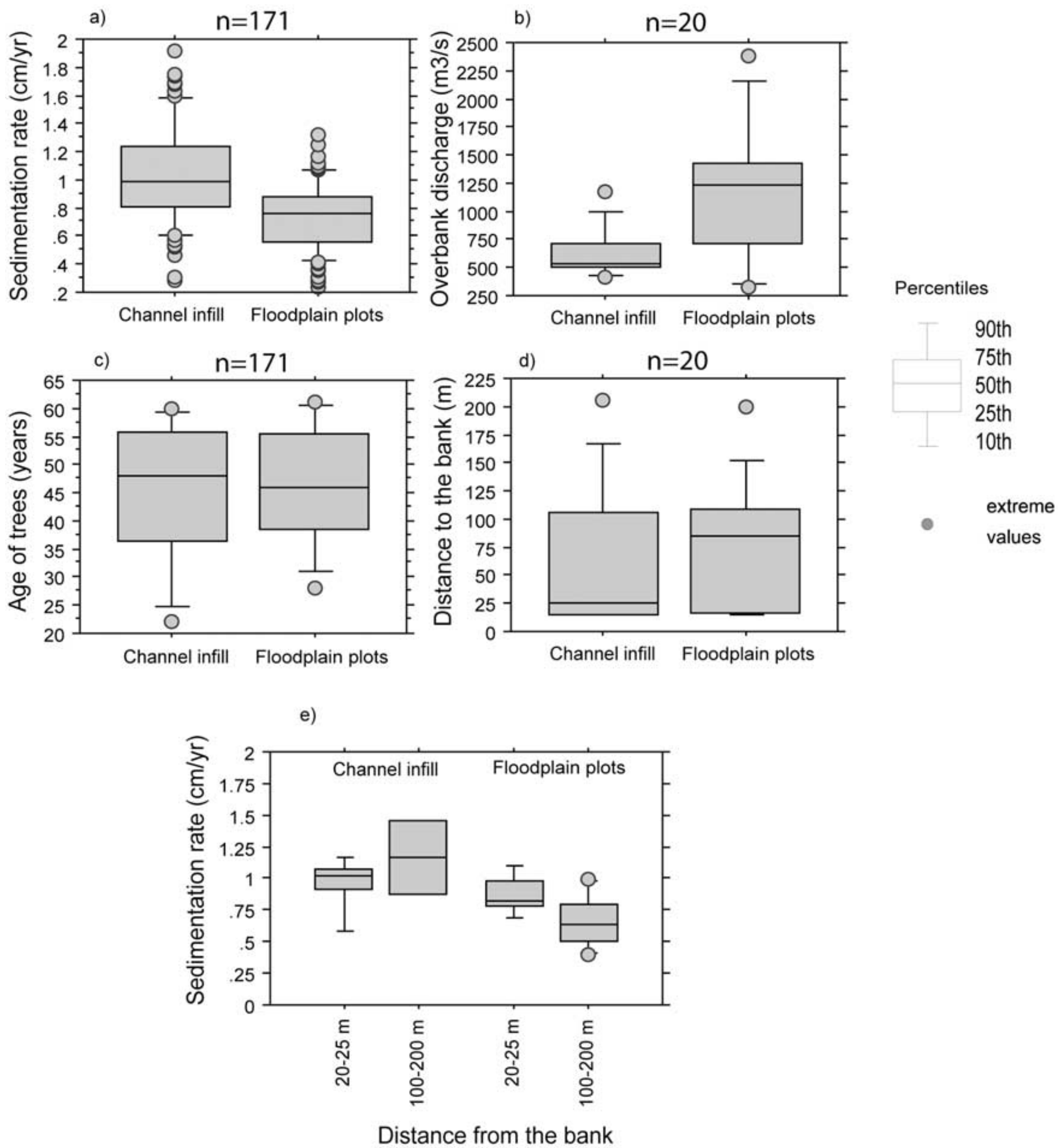

Figure 8. Comparison of sedimentation rate and local characters between the channel infills and the alluvial forest plots: Box plots of (a) the sedimentation rate, (b) the overbank flow discharge, (c) the age of trees, and (d) the distance to the bank in the channel infills and the floodplain plots and (e) the sedimentation rate in channel infill and floodplain plots according to the distance to the bank. Here n is the number of trees sampled for each of the groups. Data not log transformed to illustrate variability.

and 1945. An overbank flow channel is incised into the floodplain surface (Figure 10a) and is hydrologically connected when the discharge exceeds 270 $m^3 s^{-1}$, whereas the adjacent forested floodplain is flooded only after the discharge exceeds 620 $m^3 s^{-1}$. The slope of the channel infill is 0.0042 (Figure 10b). The geometry of the abandoned channel is difficult to detect along its upstream length from the topography and also from the position of the gravel layer (Figure 10c). However, the gravel layer position appears in cross section 4 (CS 4) and in CS 7/CS 8 and a well preserved former channel is apparent in CS 10 and CS 11. This channel conducts groundwater at CS 6 during low-flow discharges. CS 9 to CS 11 support the upstream part of the lake, with macrophytes on the left side. The main channel bank eroded on average 14 m a^{-1} between 1980 and 2000 (total distance: 170 m), a 37% decrease in channel infill length (Figure 2d). Sediment cores were taken along the overbank flow channel and along the lake on CS 10 and CS 11. Six plots where also sampled for tree age (p1 to p6 on Figure 10a). It was possible to core trees in age cohorts within each plot for temporal analyses; ages ranged from 12 to 60 years.

[38] The sedimentation rate gradient from the bank to the lake is relatively well defined, decreasing from 1.5 cm a^{-1} near the bank to 0.25 cm a^{-1} near the lake. Most of the plots show a clear gradient of sedimentation from the youngest

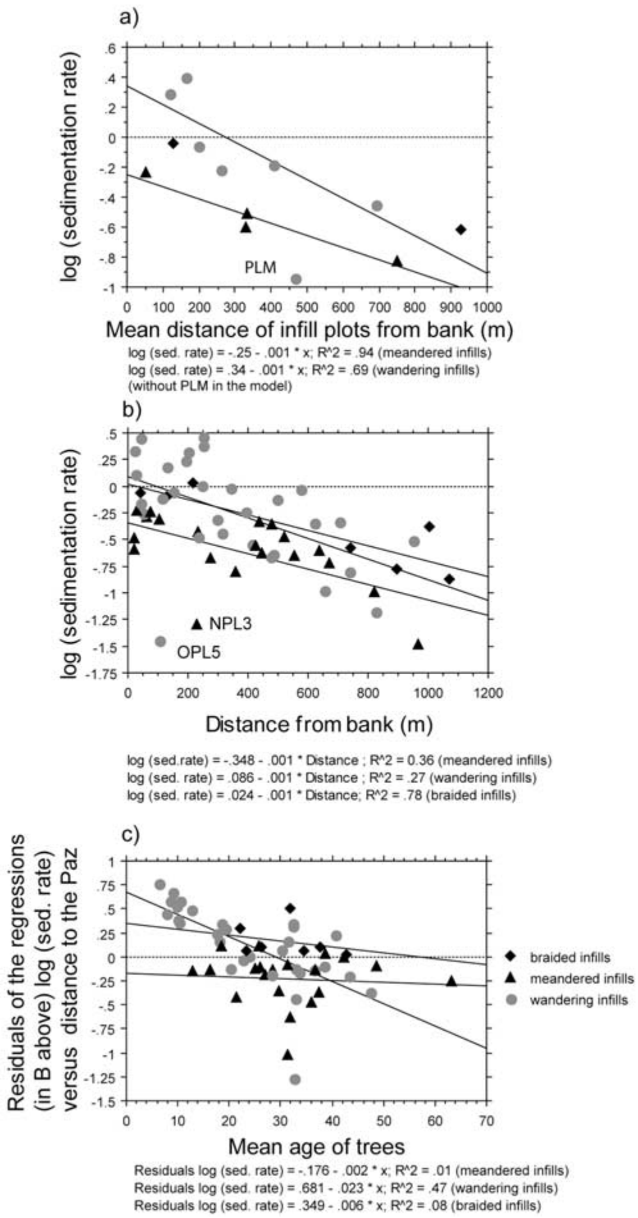

Figure 9. Decrease in log-transformed sedimentation rates from the bank to the lake according to the channel infill types (a) using mean values provided for each channel infill and (b) using mean values provided for each plot within each of the channel infills (OPL5 means OPL station plot 5). (c) Residuals of the models established in Figure 9b for each of the channel infill types.

trees to the oldest with tree age inversely related to sedimentation rate (Figure 11). Over the last 40 years, the sedimentation rate increased from 0.13 to 0.45 cm a^{-1}, with the highest increase occurring in plots p1, p2 and p5. This increase is less clear in plots p3 and p4, the latter having no trees younger than 30 years. Plot p6 located near the channel bank does not follow this trend, having a high sedimentation rate regardless of the age of the trees and a high variability between trees of similar ages. On plots p1, p2, p3, and p5, a break occurs in the sedimentation rate 25 years ago, a significant difference in sedimentation rate is therefore observed between the last 25 year period and the previous period (Mann-Whitney test: $z = -2.661$; $p = 0.078$). Before about 1975, it seems that the sedimentation rate was fairly constant through time but then underwent a continuous increase from 0.25 to 0.5 cm a^{-1} on plot p1, and 0.5 to 1.75 cm a^{-1} on plot p5.

[39] The radionuclide analysis shows that, with the exception of core CS 11, the estimates of the total ^{137}Cs inventory obtained for the cores are significantly greater than the local reference inventory, thereby providing clear evidence of sedimentation (Table 1). Comparison of the longitudinal variation of the sedimentation rates estimated from the 1963 ^{137}Cs peak within the ^{137}Cs depth profiles (Figures 12 and 13) with that exhibited by the dendrogeomorphic data, provides evidence of a similar and consistent trend. There is a clear decrease of sedimentation rate from the bank to the lake. The site closest to the lake (CS 11), provides no evidence of sedimentation, since its total ^{137}Cs inventory is less than the local reference value. The well defined exponential depth distributions found for both ^{137}Cs and ^{210}Pb at CS 11 (Figure 12) provide clear confirmation of a lack of significant mixing in the surface sediments of the deposits and this adds further confidence to the identification of the 1963 surface within the other cores.

[40] The dendrogeomorphic data are validated by the radionuclide activity profiles, which vary significantly from the bank to the lake, demonstrating that the sedimentation history is not similar from the upstream to the downstream parts of the infill deposit. The radionuclide profiles (Figure 12) show that cores CS 1 and CS 4 are characterized by different shapes than cores CS 6 and CS 8. Cores CS 9 and CS 11 also exhibit different shapes. The unsupported ^{210}Pb depth profiles for all cores except that for CS 11 provide evidence of two concentration peaks, with one at the surface and the other in the middle part of the core, which may reflect temporal variation in the sedimentation rate and/or changes the particle size composition of the deposited sediment (Figure 12).

[41] The shapes of three of the ^{210}Pb profiles (1, 4 and 9) depart significantly from the normally expected progressive down core reduction in activity. Cores from CS 1 and CS 4 are characterized by a 10 cm deep surface layer with a lower ^{210}Pb content than the deeper sediment and this can be tentatively interpreted to reflect an input of coarser deposits with lower ^{210}Pb activity, related to recent higher-magnitude flood events. Sediment deposits associated with large flood events may contain lower fallout radionuclide activities than sediment associated with smaller flood events, due to an increase in coarse particle content and a shift in sediment sources. Thus, burial of a sediment surface containing high levels of unsupported ^{210}Pb by a thick layer of sediment containing low levels of unsupported ^{210}Pb and the subsequent incorporation of atmospheric ^{210}Pb fallout into the new surface layer could result in unsupported ^{210}Pb profiles that are similar to those observed in our sediment cores. Contrasts in profile shape between the individual cores may also reflect the impact of bank erosion in progressively reducing the distance from coring plots to the channel bank and thus changing the input of overbank sediment along the infill. Such an increase in sediment delivery due to the change in channel position is likely to have particularly affected cores CS 1, CS 4 and CS 9. The locations of cores

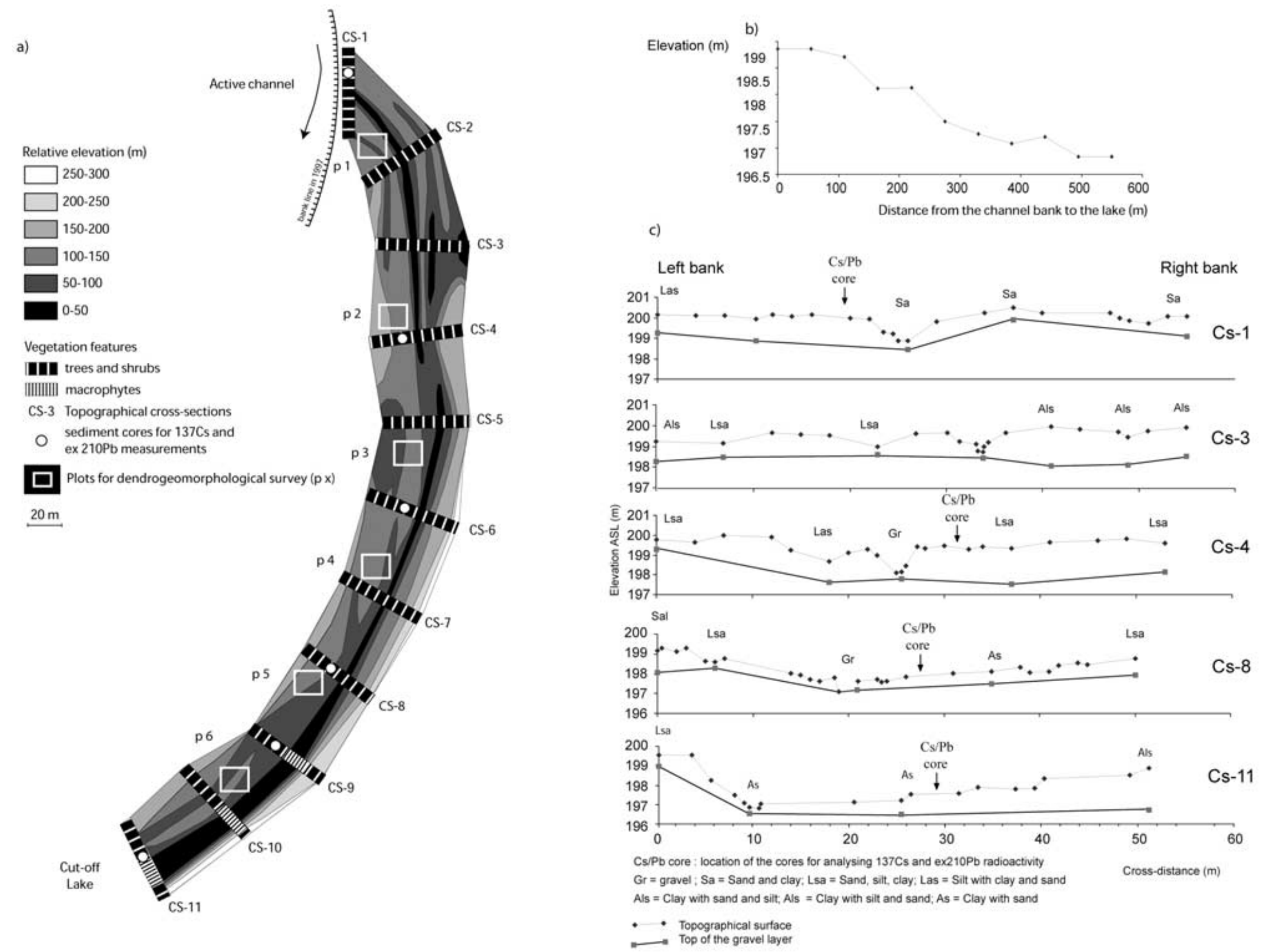

Figure 10. General characteristics of the RIC cutoff channel infill: (a) location of the dendrochronological plots and the sediment cores from which sedimentation rate has been estimated from ^{137}Cs measurements, (b) long profile of the channel infill from minimum elevations at cross sections, and (c) examples of cross sections with the field topography and the top of the gravel layer.

CS 6 and CS 8 reflect somewhat different conditions. These sites are shielded from the channel by a wide forest area, and no significant change in sedimentation is likely to have occurred over the past decades.

5. Discussion

5.1. Comparison of the Sedimentation Rate of the Ain With Other Rivers

[42] With a range of floodplain sedimentation rates varying between 0.65 and 2.4 cm a^{-1} over a period of 10 to 40 years, the lower Ain River is typical of French piedmont rivers, with similar values reported for the Garonne River near Toulouse (e.g., 0.5 to 2.5 cm a^{-1} [*Steiger et al.*, 2001b]). These rates are higher than along most lowland floodplains of Coastal Plain rivers in the USA [*Mitsch et al.*, 1979; *Cooper et al.*, 1987] where rates typically range between 0.15 to 0.54 cm a^{-1} [*Hupp*, 2000] or along the lower Rhine River [*Middlekoop*, 2002] where rates range from 0.2 mm to 15 mm a^{-1}. High rates are also observed along rivers which have undergone a period of very high suspended sediment transport, such as the Tyne Valley, England, with 5 to 7 cm a^{-1} between 1890 and 1930, which is exceptionally high for the British standard [*Macklin et al.*, 1992] and related to mining activity. The Isère River in the northern French Alps has experienced local sedimentation rates of up to 85 cm a^{-1} related to a highly erosive geology within catchment [*Peiry*, 1997].

5.2. Sedimentation Rate and Local Overbank Discharge Frequency

[43] Our results provide additional support regarding the effect of local overbank discharge frequency in controlling sedimentation rates in floodplains. These results underscore the unique character of channel infills compared to the typical forested floodplain in terms of sedimentation rate. Because they are lower, they undergo more frequent inundation and hence exhibit a higher sedimentation rate.

[44] This relationship, however, also depends on the infill types, where wandering infills experience higher sedimentation rates than braided or meandering channel infills for a given discharge. The wandering channel infills are not significantly longer or more frequently flooded than the other channel infills, but the geometry of the former

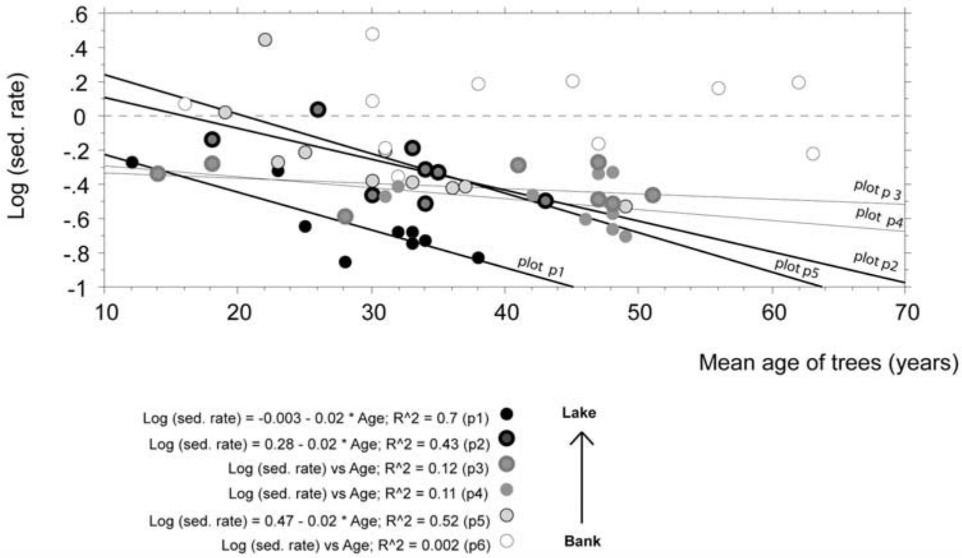

Figure 11. Trend in log-transformed sedimentation rates versus the tree ages for each of the sampling plots of Ricotti (RIC), distributed from the bank to the lake.

channels may have a substantial effect, with entrance angle θ relative to the main channel varying from a few degrees for most of the wandering infills to about 90° for some meandering channels. The lowest angle occurs when the wandering is located on the axis of outside bends (cut-banks). These highly connected channel infills are closer to the channel on average and may trap more sediment for a given flood frequency.

5.3. Distance to the Bank: Suspended Sediment Deposits Versus Bed Load Sand Deposits

[45] A clear link exists between sedimentation rate and distance to the channel, as shown by other authors in the floodplain context [*Walling and He*, 1998; *Ross et al.*, 2004]. It is highlighted in the floodplain plots of the Ain, and along some of the channel infills.

[46] This longitudinal trend is most distinct where the channel infill has maintained a relatively pristine channel form and substantially conveys flow during overbank events. When the angle θ is low, the main channel usually is much closer to the channel infill and can provide overbank sediment all along the channel infill without any gradient from the bank to the lake. Thus, some wandering infills are characterized by near-lake sedimentation rates higher than those near banks (Figure 6d).

[47] Also, channel infills concentrate flow and have critical shear stress farther inside the floodplain so that they can transport bed load sands through the infill up to the lake. In such conditions, they exhibit high sedimentation rates even 100 to 200 m from the banks, and significantly higher than that observed on neighboring floodplain plots. This scenario is similar to that described for levee crevasse/splay systems on the Coastal Plain of southeastern U.S. [*Hupp*, 2000, *Ross et al.*, 2004].

5.4. Change in Time Due to Changes in Local Overbank Flow Frequency

[48] Previous research has shown that a change or regime shift in sedimentation dynamics may occur, notably resulting from land use changes after European settlement in Eastern U.S. [*Jacobson and Coleman*, 1986; *Lecce*, 1997] or New Zealand [*Gomez et al.*, 1998] or post World War II changes in agricultural and pastoral practices [*Liébault and Piégay*, 2002; *Walling et al.*, 2003; *Piégay et al.*, 2004] but also from global oscillation cycles [*Aalto et al.*, 2003]. Along the Ain River, channel degradation, aggradation, and bank erosion (that affects the distance between the channel and the lake) play a role in changing sedimentation rates and creating a complex mosaic of sedimentation within the floodplain (see Figures 3 and 5). Channel instability at the reach scale, particularly degradation and aggradation processes, substantially affect the length and duration of flooding of the cutoff channel infill, local sediment deposition/erosion patterns, and the scouring potential in cutoff lakes [*Bornette et al.*, 1996; *Bravard et al.*, 1997; *Piégay et al.*, 2000]. *Wyzga* [2001] previously hypothesized possible reductions in sedimentation rates along southern Polish rivers that experienced channel degradation and a subsequent reduction in frequency and magnitude of floodplain flooding. Such changes can occur through changes in channel-floodplain connectivity as shown by our results, which demonstrate that channel

Table 1. Radionuclide Inventories Associated With the Bulk Soil Reference Cores and the Channel Infill Cores[a]

	^{137}Cs, Bq/m^2	^{210}Pb, Bq/m^2
B[b]	2100	2570
MG[b]	2430	3200
ML[b]	3040	4100
CS-1	6070	2400
CS-4	6910	4590
CS-6	5380	3660
CS-8	7420	6330
CS-9	6930	6050
CS-11	1620	1960

[a]MG, marshland of Giron, in right side of the river at Chazey; ML, Morte aux Loups, southeast of SSB; B, Blyes, near the sewage treatment works.
[b]Nonflooded areas (referenced bulk samples) within the valley bottom.

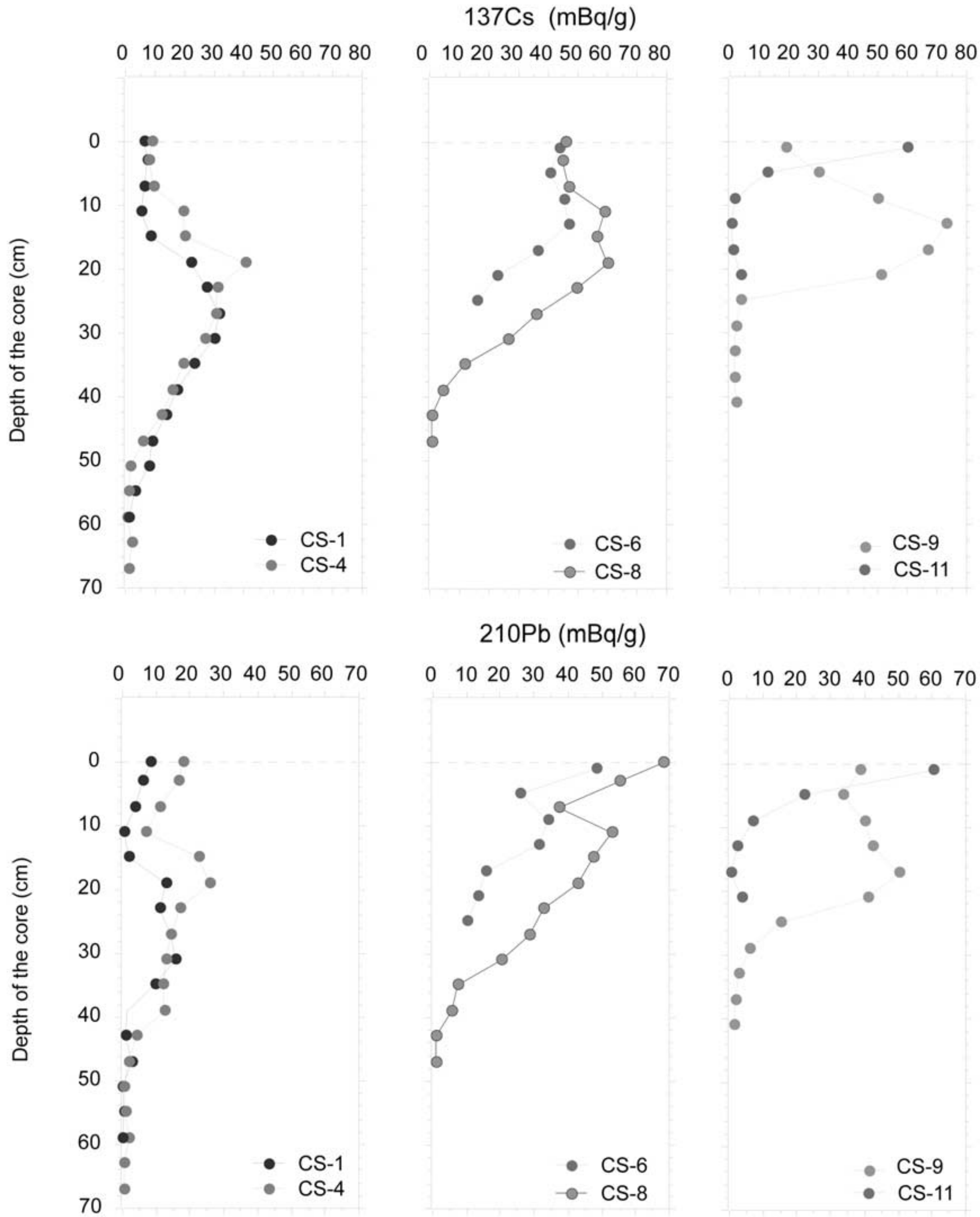

Figure 12. The ^{137}Cs and ^{210}Pb profiles from the cores of RIC from the bank (CS-1) to the lake (CS-11; see Figure 10 for location).

degradation affects flooding frequency and reduces sedimentation rates. The greatest wandering infill sedimentation rates occur along reaches that were the least degraded over decadal timescales (Figures 5d and 5e).

[49] The lateral channel movement is also an important factor that affects temporal sedimentation rate. Figure 7, shows that channel movement away from the channel infill can decrease connection from the active channel and reduce sedimentation rate, whereas the narrowing infill due to bank erosion can also favor an increase in local overbank flow frequency and associated increase in sedimentation rate. The Ricotti channel infill is shortening in length through time responding to substantial upstream bank erosion. Thus, active channel dynamics affect channel infills, which were previously less connected from the main channel and associated sediment delivery. This phenomenon explains the complex pattern of sedimentation from the bank to the lake as shown by the ^{210}Pb profiles. A recent sand wave delivered by the channel and observed to be burying the upper channel infill as far as CS 5, explain the ^{210}Pb profiles of CS 1 and CS 4 with coarser sediment causing a reduction in ^{210}Pb activity at the top of the core. Downstream, the

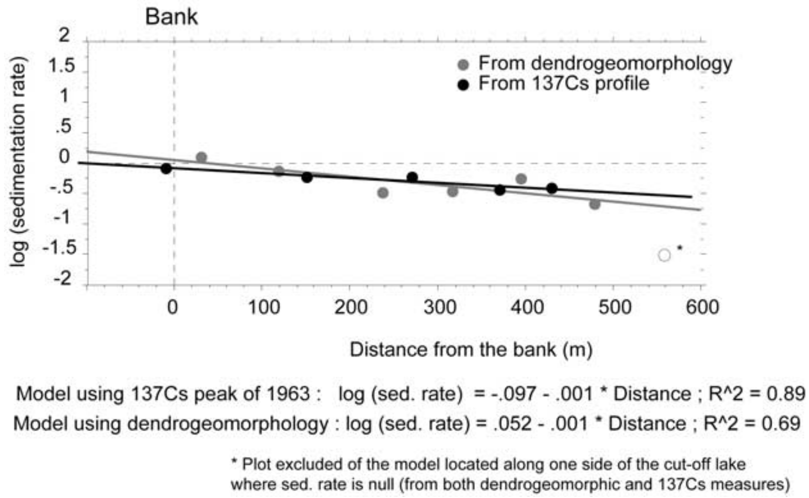

Figure 13. Decrease in log-transformed sedimentation rates from the bank to the lake according to the dendrochronological analysis and to the ^{137}Cs calculation.

flow and the sediment flux are concentrated in the overbank channel, which is clearly drawn in the channel infill morphology as far as CS 9. The sand is then deposited in the lake, which appears at CS 9 resembling an alluvial fan that fills the upper part of the lake, and may explain the shape of the depth profile of ^{210}Pb found in core CS 9. CS 11 remains unaffected by the reactivation of the area and the recent influx of sand, which suggests a limited effect of floodplain rejuvenation on the lower part of the channel infill and the immediate margins of the overbank channel. Almost no sedimentation occurs at CS 11, particularly over the past 25 years. These results therefore demonstrate that sedimentation rates may vary through time and not systematically as *Hooke* [1995] has shown with a reduction in deposition rate related to a decrease in frequency of overbank flooding.

[50] We can, nevertheless, postulate how this complex functioning system will evolve through time and determine if similar situations exist elsewhere. It is reasonable to speculate that the Ain River is adjusting to new conditions, while the previous braided and wandering patterns still affect a landscape fingerprint with remnant cutoff channels. Future conditions should support much less diversity in overbank sedimentation patterns within the floodplain when the remnant features become rejuvenated by the, now, meandering channel shifting. The river may be approaching a "temporal ecotone" in the sense described by *Naiman and Décamps* [1990] where the historical wandering features are located in specific areas of the fluvial corridor because of the way the channel narrowed and progressively increased its sinuosity (e.g., they are mainly behind the actual active concave meander banks undergoing advective sand inputs with a low angle to the main channel). Whereas the braided infills are more isolated from the meander band as well as the meander infills, which have a high angle in respect to the main channel.

5.5. Technique Accuracy and Constraints

[51] Techniques used to establish floodplain sedimentation rates are quite varied, as shown by *Steiger et al.* [2003] in relation to the timescale considered. Nevertheless, the most common approaches for analyzing periods of 1 to 10 decades are the use of short period radionuclides (^{137}Cs and unsupported ^{210}Pb) and dendrogeomorphic surveys. However, both techniques are infrequently used in combination. Thus, our study is unusual in this regard in that we cross validated the dendrogeomorphic approach with the radionuclide approach, both of which are independent and sensitive to different limitations and sources of error.

[52] Although dendrochronological timescales may provide useful information, they have constraints when used in studies of fluvial surface initiation and development. The age of the cutoffs is linked to the maximum age of the trees, previously demonstrated by others where trees may establish shortly after the formation of a "new" riparian surface [*Nanson and Beach*, 1977; *McKenney et al.*, 1995; *Friedman et al.*, 1996]. The sedimentation rate is strongly associated with the age of the surface [*Hupp*, 1988] and its progressive disconnection from the main channel in relation to floodplain accretion [*Nanson and Beach*, 1977; *Scott et al.*, 1997]. The dendrogeomorphic analysis then yields a good estimate of the mean net sedimentation rate over the life of the channel infill.

[53] In this study, it has been shown that tree ages are independent of surface age. When tree ages are different from surface ages, (because of human activity such as clear cutting, e.g., BUM or because the observed tree ages correspond to a well established stage of the ecological succession, e.g., *Fraxinus excelsior,* the preferred tree for sampling but not usually among the first colonizers), age data may over or under estimate changes in sedimentation rate in comparison to sedimentation rates observed over the period of surface development. Thus, it is important to know the time period captured by the dendrogeomorphic survey to interpret any change in sedimentation related to the entire evolution of the surface.

[54] Trees that establish early in the evolution of fluvial surfaces may clearly reflect the true temporal trends, which is the case at the BUM site where tree ages are young and sedimentation rates are high (Figure 7c). If the period covered by the dendrogeomorphic survey is based on old trees as at VIL and PLM, recent sedimentation rates may be

underestimated. VIL and PLM surfaces have the oldest tree specimens (e.g., 40 years old, Figure 5d) and the sedimentation rate recorded on the two sites did not capture short-term trends, associated with a possible change in connectivity due to channel aggradation between 1976 and 1999. Nevertheless, these two channel infills are located at a high elevation (local overbank discharge for VIL is 1200 $m^3 s^{-1}$) and are rarely affected by overbank flows (floods > Q_{10}). Thus, recent channel aggradation may not substantially influence changes in sedimentation rates along adjacent channel infills. Figure 9c shows that the residuals of the regression between the sedimentation rate and the distance from the bank are linked with the mean age of trees. The younger the trees are, the more positive the residuals of the regression, whereas some wandering channel infills are quite old (Figure 7a). Typically, the tree age does not correspond to the age of the channel infill surfaces but provides an estimate of the sedimentation rate for recent periods of time, suggesting that sedimentation rates tend to increase in recent periods compared to those previous, mainly in response to increased flooding frequency, because these channel infills are located downstream of cutbanks and underwent continuous regeneration (see Figure 3).

[55] In a retrospective study, the time periods covered by each of the parameters (e.g., the potential control parameters—channel movement, aggradation versus degradation, flood history—but also the target parameters, floodplain sedimentation rate in this context) are an important consideration that may affect interpretation. Each parameter may have a site-specific time coverage that is dependent on available sources of historical information (tree ages or historical documents, which do not match the timing of major events in the flood series or breaks in the geomorphic trends). In such context, a site that has different tree-age cohorts (e.g., Ricotti) or at least trees of various ages, can substantially benefit from the characterization of trends in sedimentation rates that leads to the identification of controlling factors that affect important changes. On the basis of a comparison with historic records, our dendrogeomorphic results appear to be the most accurate when the annual sedimentation rate is not exceptionally high, between 0.1 to 10 mm a^{-1}. When a site undergoes several centimeters of sedimentation per year over a few decades (e.g., recent deposits along the banks of Ricotti), the sedimentation estimates are more difficult to establish dendrogeomorphically with a few samples. Furthermore, it is difficult to accurately determine the correct depth of burial without an extensive excavation of the root system. The presence of adventitious roots above the original root collar also complicates efforts and may lead to underestimation of sedimentation rates. These limitations may explain why no trend in sedimentation rate could be established on the upstream plot at Ricotti (RIC). Here a limited number of trees younger than 30 years prevented the determination of recent sedimentation trends (plots 1, 2 or 5; see Figure 11).

[56] The use of the ^{137}Cs peak to estimate sedimentation rates can be affected by grain size compositional changes along the core. Because ^{137}Cs is preferentially associated with finer fractions, such compositional changes may affect the ^{137}Cs profile. The ^{137}Cs profiles usually show a concentration peak in the middle part of the profile with a decline in concentration both toward the soil surface and below the peak. Although the profiles documented in this study are similar to those observed in other river floodplains, they also have significant differences, in terms of both ^{137}Cs profile shape and total inventory in relation to grain size variability. Overbank sediment deposits generally become finer as the distance from the river channel increases, which can be expected to result in higher ^{137}Cs and unsupported ^{210}Pb concentrations in sediment deposited farther from the channel than in sediment deposited closer to the channel. Both ^{137}Cs and fallout ^{210}Pb are primarily associated with fine particles. This behavior of overbank floodplain sedimentation and the absorption of fallout radionuclides by soil particles may partly explain the lower ^{137}Cs and unsupported ^{210}Pb concentrations associated with the sediment from the upper part of the CS 1 and CS 4 cores. In the case of important change in sedimentation pattern, the ^{137}Cs profile may be affected and the 1963 peak may not be clearly observed, as shown in Figure 12. Nevertheless, the profile shapes are judged to be sufficiently robust to accurately estimate an average sedimentation rate over the past 4 decades. It is therefore compelling to use ^{137}Cs and ^{210}Pb profiles simultaneously, as they provide complementary information that facilitates interpretation of complex features such as radionuclide profiles in changing environments (e.g., the floodplains of Alpine systems [see also *Piégay et al.*, 2004]).

6. Conclusions

[57] Sedimentation rates in cutoff channel infills have a similar pattern to those previously observed in floodplains. They usually decrease in space from the proximal zone (the bank) to the distal zone (the lake), but also in time as the overbank flows are rare. But cutoff channel infills also undergo higher sedimentation rate than the floodplain because they are more frequently flooded. Because the active infills can convey bed load sediment to the lake, especially the wandering types with a low θ angle, they do not systematically exhibit a clear decrease in sedimentation rate along their course. Moreover, these sediment patterns are not constant through time. Because of channel movement that narrows or lengthens the infill deposits, but also channel degradation or aggradation modifying the magnitude and frequency of local overbank flow, sedimentation rate increases or decreases through time. This suggests that the geomorphic processes occurring over relatively short timescales, such as tree ages (e.g., 40 years), are different than those associated with the creation and life span of channel infills. At this long-term scale, channel infill sedimentation rates may have evolved differently from one channel infill to another.

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