PERGAMON

Quaternary Science Reviews 22 (2003) 2061–2076

QSR

# Response of the Selle River to climatic modifications during the Lateglacial and Early Holocene (Somme Basin-Northern France)

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Received 24 January 2002; accepted 22 April 2003

**Abstract**

Research on Lateglacial sequences from the Selle valley leads to an overview of its evolution in relation to climatic variations between the end of the Weichselian Upper Pleniglacial and the beginning of the Holocene. The first major modification of the fluvial morphology is dated at the Upper Pleniglacial/Lateglacial transition (13,000 $^{14}$C-yr BP). At that time, the response to climatic improvement and environmental modifications is marked by downcutting and evolution from a braided river to a transitional river pattern (Bølling infilling in the newly created channels). After a short cold phase recorded in a thin calcareous bed at the top of the Bølling peat attributed to the Older Dryas (Dr. II), the Allerød is characterized by the deposition of organic overbank silts within a large single channel meandering system. In lower slope environments, this period is also marked by slow rates of colluvial accumulation and by the development of upbuilding soils (Allerød soil). On the other hand, the end of the Lateglacial, is characterised by the infilling of the whole valley by fine calcareous overbank silts during the Younger Dryas cold phase (overflow of a large single channel and lateral input of chalk mud). A second major downcutting phase occurs at the beginning of the Holocene at around 10,000 BP, in parallel with another rapid climatic improvement and the renewed spread of vegetation. From a general point of view, the evolution of fluvial environments in the Selle valley is comparable with many other river valleys in NW Europe, showing that fluvial systems react very quickly to climatic variations of short duration (1000 to 100 years). Finally, in the Upper Selle River, incision events occur clearly before the main modifications of the vegetal cover. They are most likely linked to a rapid shift in the balance between water discharge and sediment supply, caused by climate modifications (shift to more temperate and oceanic conditions), and the resulting environmental changes: cessation in aeolian sedimentation, strong reduction of slope processes, permafrost disappearance and soil development.



## 1. Introduction

Research on the Lateglacial and early Holocene environments of the Somme Basin (Fig. 1) has developed over the last few years, primarily as a result of archaeological excavations in valley bottom deposits and at the base of hillslopes (Fagnart and Coudret, 1995, 2000; Fagnart, 1997a, b; Antoine, 1997a–c; Ducrocq, 1999; Antoine et al., 2000; Limondin-Lozouet and Antoine, 2001). This work has recently made it possible to reconstruct the Lateglacial to Holocene succession of the Somme Basin and to compare it with the sequences from other north-west European rivers (Vandenberghe et al., 1994; Tebbens et al., 1999; van Huissteden and Kasse, 2001), as well as results from the Paris Basin (Lefèvre et al., 1993; Leroyer et al., 1997; Pastre et al., 1997, 2000; Antoine et al., 2000; Limondin-Lozouet et al., 2002).

This multi-disciplinary research has brought together the work of various specialists: palynology (A.V. Munaut); malacology (N. Limondin-Lozouet); coleoptera (P. Ponel); Final Palaeolithic (J.P. Fagnart and P. Coudret); Mesolithic (T. Ducrocq). In addition,

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doi:10.1016/S0277-3791(03)00180-X

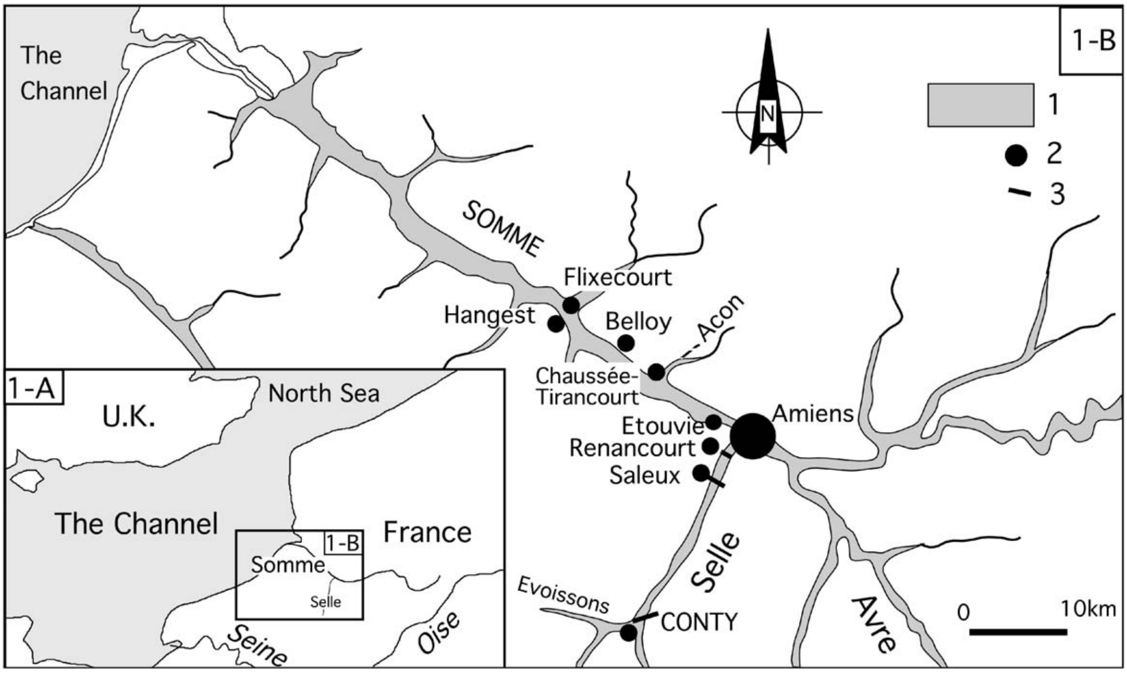



Fig. 1. (A) Location map of the Somme Basin in NW Europe. (B) Location of the cross-sections in the Selle and of some of the studied sites in the Somme Basin ((1) current alluvial plain, (2) studied sites, (3) main cross-sections of Conty, Saleux and Renancourt).

these studies have led to recognition of the interactions between human settlement and environment during the Lateglacial (Antoine et al., 2000).

Within the context of this research, the sequence at Conty was discovered in 1995, yielding an exceptional Lateglacial record that provides much new data on the pattern of evolution of valleys previously initiated in the Somme (Antoine, 1997a). The Conty site is located in the Upper valley of the Selle (discharge $\pm 5\,m^3/s$, drainage basin: $524\,km^2$), which is a small tributary entering the Upper reaches of the Somme (discharge $\pm 35\,m^3/s$, drainage basin: $5560\,km^2$). The Conty gravel pit is located approximately 20 km S-SW of Amiens, at the junction between the Selle and Évoissons valleys (Fig. 1B). The local setting is characterized by steep chalky slopes on the west bank of the valley, while the gently sloping east bank has only a thin Quaternary cover (relicts of terrace gravels and colluvium on chalk). The stratigraphic study of the valley filling at Conty is based on a detailed transect (600 m, 35 borings, and 8 core samples taken using PVC tubes: $16\,cm \times 100\,cm$, Fig. 2). Other sites have been studied farther downstream in the Selle valley at Saleux, during excavations of an important Final Palaeolithic site (Fagnart and Coudret, 1995), and at Renancourt, close to the confluence with the Somme, during the prospection of Final Palaeolithic and Mesolithic sites (Ducrocq and Antoine, 1996) (Fig. 1B).

The synthesis of the stratigraphic, geochronological and environmental data obtained over the whole of the Selle valley leads to a tentative reconstruction of the impact of climatic variations on fluvial morphology and sedimentation, as well as an attempt to quantify the sedimentary budgets during the principal phases of evolution of the valley.

## 2. Weichselian pleniglacial background

In the Somme basin, as in the whole of northern France, the end of the Pleniglacial is characterized by a wide extension of loessic deposits (Lautridou and Sommé, 1974; Sommé et al., 1980; Lautridou, 1985; Antoine et al., 1998, 2000). This Upper Pleniglacial loess blanket was deposited as a continuous cover on the plateaus and slopes (“Cover Loess”). Lithostratigraphic correlations on the scale of north-west Europe, using marker horizons such as the Kesselt/Nagelbeek tongue Horizon (Haesaerts et al., 1981; Juvigné et al., 1996), and the new TL-IRSL ages obtained in the Somme (Engelmann and Frechen, 1998), yield dates for these last phases of loess accumulation between ca. 22–20 and 16–15 kyr BP (Antoine et al., 1999).

At the edges of alluvial sheets on the lower terrace (Montières and Étouvie Formations, respectively located at +5–6 m and +10–12 m relative height above the maximum incision; Antoine, 1990), or at the bottom of slopes near the river valley, these typical calcareous loess deposits pass laterally and progressively into more heterogeneous loess, characterized by the presence of chalk granules, sand grains and small frost-shattered flint fragments (Antoine, 1997a). This lateral facies

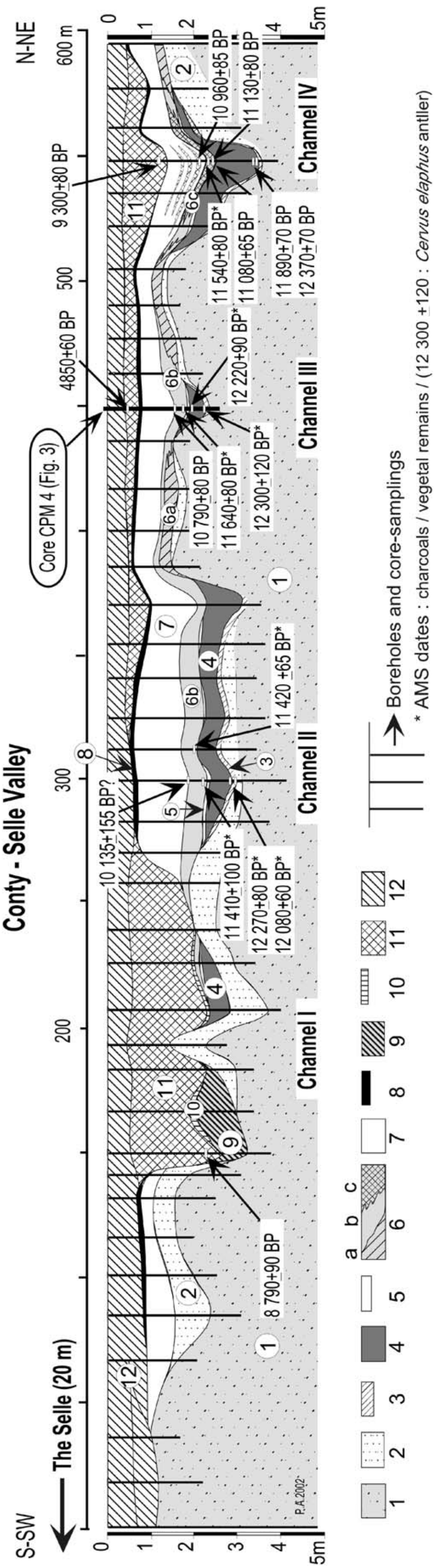



Fig. 2. Detailed cross-section through the Selle valley at Conty (modified after Antoine, 1997b). (1) valley-floor gravel, (2) Loessic silt, (3) slightly organic silt, (4) "reddish peat" with numerous plant remains (wood, bark, seeds) and molluscs, (5) thin horizon of white calcareous silt, (6a) calcareous organic silts with chalk granules, (6b) grey organic silts with organic horizons and final palaeolithic artefacts, (6c) lateral peaty facies of (6b), (7) homogeneous and light-grey silt with abundant terrestrial molluscs and organic horizons at the base containing aquatic molluscs, (8) clayey grey-black organic silt, (9) flint gravel with organic matrix and plant remains, (10) granular calcareous tufa (lens at the bottom of channel I), (11) black peat and (12) complex of clayey silts with Gallo-Roman level covered by the current valley soil.

variation results from the mixing of the allochtonous loessic particles with local coarse material, derived from deflation at the top of sandy-calcareous fluvial deposits of the Pleniglacial braided river plains (chalk granules, sand, little flint fragments). In the Selle River, as in the Somme, the Upper Pleniglacial loess extends widely onto the alluvial plain, where it overlies most of the area covered by fluvial gravels (loessic filling of the valley at the end of the Upper Pleniglacial) (Fig. 2, unit no. 2).

The loess is generally barren or very poor in pollen and molluscs, suggesting a cold steppe to polar desert setting contemporary with the coldest and driest phase of the Pleniglacial (Marine Isotope Stage 2). Moreover, during this very arid period at the end of the Upper Pleniglacial, the valleys (whatever their size), were occupied by a system of braided channels separated by gravely bars, typical of many rivers in north-west Europe at that time (Vandenberghe et al., 1994; Mol, 1997; Huijzer and Vandenberghe, 1998; Huisink, 2000; Van Huissteden and Kasse, 2001). These structures reflect a strongly seasonal discharge regime with primarily active dynamics during the breaking up of ice in spring.

In such a cold-climate environment, where the permafrost prevents deep infiltration, mobilization by gelifluction of the active layer on the slopes provides an abundant supply of coarse sediments (chalky mudflows with flints) which block up the alluvial plain (exceeding the transport capacity of the river).

While the heterogeneous loess sequences on the slopes and higher parts of the valley are systematically unfossiliferous, the upper part of the hydromorphic silts in the valley bottom locally yield a rather rich malacological fauna with *Vallonia costata* and *Vallonia pulchella*. The faunas led Limondin (1995) to attribute these deposits to an early Lateglacial stage because abundant populations of these species are unable to develop in cold Pleniglacial environments and have never been found in typical Upper Pleniglacial loess. Palynological data from the base of boring S18 in the Acon valley (Fig. 1B) also seem to indicate an early Lateglacial age for the rather thin (5–15 cm) silty bed located at the bottom of the core (Lambay, 1993). The recent malacological study of the Conty site shows that these "*Vallonia faunas*" are attributable to the biozone Cty-1 at Conty, representing the base of the Bølling (Limondin-Lozouet and Antoine, 2001).

Finally, in this paper, the lower boundary of the Lateglacial (Fig. 5) is correlated with the base of the Bølling oscillation (Antoine, 1997a, b), according to an increasingly adopted view (Walker et al., 1994; Mol, 1997; Huisink, 1998, 2000; Van Huissteden and Kasse, 2001). Indeed, since it remains very difficult to detect the Dryas I (Oldest Dryas) event in the stratigraphic record, the major break at the base of the Bølling (change in the fluvial style and incision processes) provides a boundary that can be identified during field-works and used for correlations on the scale of the basin.

Finally, to avoid problems of correlation between pollen-based units (e.g. discussion about the Meiendorf and the Bølling in Litt et al., 2001), we use the term Bølling in its chronostratigraphic sense to describe the first major Interstadial of the Lateglacial dated between 12 and 13 $^{14}$C-kyr BP (about 14 to 14.7 kyr cal. BP). This Bølling chronozone thus corresponds to the GI-1e event of the INTIMATE Group event stratigraphy (Björck et al., 1998; Walker et al., 1999).

## 3. Principal phases of modification in valley-bottom environments during Lateglacial and early Holocene times

### 3.1. Bølling: concentration of channels, initial downcutting then localised peaty filling

The reference transect of the Selle valley at Conty (Fig. 2) records the transition of a periglacial system (with braided channels) towards a transitional system characterised by several low sinuosity narrow channels (according to the definition of Vandenberghe, 1993; Vandenberghe et al., 1994 and Kasse et al., 2000).

These modification is accompanied by a major phase of incision (2–2.5 m) into the Pleniglacial deposits (gravels, loess) in the early Lateglacial (Antoine, 1997a, b). According to $^{14}$C dating results (AMS on wood remains, hart antler, and bulk samples), obtained from the basal part of the peaty fillings of these channels (Figs. 2 and 3, base of unit no. 4), the downcutting appears to have taken place at the beginning of the Bølling event, before 12,400 $^{14}$C-yr BP: 12,370 $\pm$ 70 BP (LY 6998) (14,827–14,150 cal. BP) and 12,300 $\pm$ 120 BP (OxA-6257) (14,874–14,036 cal. BP). (average age: 14,472 cal. BP, Fig. 5).

We should stress that the age of the oldest organic infilling of the channels at Conty is very close to that obtained in the Meuse River by Tebbens et al. (1999) (12,330 $\pm$ 120 $^{14}$C-yr BP).

In the Selle River, this evolution of the regime and fluvial morphology takes place gradually through a transitional system, characterized by the presence of stable, well-individualised and slightly asymmetrical channels (Fig. 2, channels II to IV, Fig. 5).

The downcutting observed in the Selle river is a typical response of river systems to the rapid climatic improvement during the Bølling, as described in most river valleys of NW Europe (Kozarski, 1983; Haesaerts, 1984; Bohncke et al., 1987; Vandenberghe et al., 1987, 1994; Kalicky, 1991; Vandenberghe, 1993; Roblin-Jouve and Rodriguez, 1997; Leroyer et al., 1997; Pastre et al., 1997; Tebbens et al., 1999).

Based on morphostratigraphic data from Conty, the sedimentary volume eroded in the Upper Selle river during the downcutting phase at the beginning of the Lateglacial can be evaluated at approximately 300 000 m$^3$ per km$^2$ of catchment area (Fig. 5). Since we cannot locate any thick deposits coeval with this phase in the middle valley of the Somme, this implies that the sediments were evacuated and probably accumulated in paleochannels of the Somme currently situated under the English Channel (Auffret et al., 1982). Such paleochannels were largely emergent at the beginning of the Lateglacial (sea level lower than $-60$ m).

The end of the Bølling phase is marked in the Conty sequences by a sudden cessation of peat accumulation in the channels, followed by deposition of calcareous fluvial silts (Fig. 3, unit no. 5, 5–10 cm thick). These sediments contain up to 55–60% of $CaCO_3$, and show a "chalk-mud" facies closely similar to that of the Younger Dryas (Fig. 2, unit no. 7).

On the basis of $^{14}$C dates obtained on encasing layers, this deposit (Figs. 2 and 3, no. 5) is attributed to the Older Dryas (Dryas II), which is compatible with its stratigraphic position and sedimentological characteristics as well as its malacological (occurrence of Arctic-Alpine snails such as Columella columella), and palynological contents (lowering of the AP/NAP ratio, Fig. 3) (Antoine et al., 2000; Limondin-Lozouet and Antoine, 2001). This climatic episode has been previously recognized from palynological evidence at Famechon a few km upstream in the Évoissons valley (Fig. 1C) (Emontspohl and Vermeersch, 1991).

Finally, assuming this chalk-mud facies was deposited as overbank deposits during a renewed phase of marked fluvial activity (new period of flooding in the valley following the peaty infilling of the channels), this event cannot be solely the consequence of a change in river dynamics. Indeed, the previously discussed biological proxies and the sedimentological characteristics of this unit (detrital $CaCO_3$: 55–60%, TOC $\leqslant$ 3%) indicate a distinctly cold climate as well as the enhanced formation of chalk debris and chalk mud by freeze-thaw processes on hill-slopes bordering the river valley (see Section 3.3). In the Selle valley the recording of the short-lived climatic cooling of the Older Dryas is clearly enhanced by the local lithology of slopes surrounding the valleys, as during the Younger Dryas.

On the other side of the English Channel at Holywell Coombe, on the same type of Upper Cretaceous chalk bedrock, a strong reactivation of slope dynamics, leading to the deposition of chalky sediments (chalk mud) during this event (O.D.), is also recorded during the same time span (11.8 to 11.5 $^{14}$C-kyr BP; Preece and Bridgland, 1998, 1999).

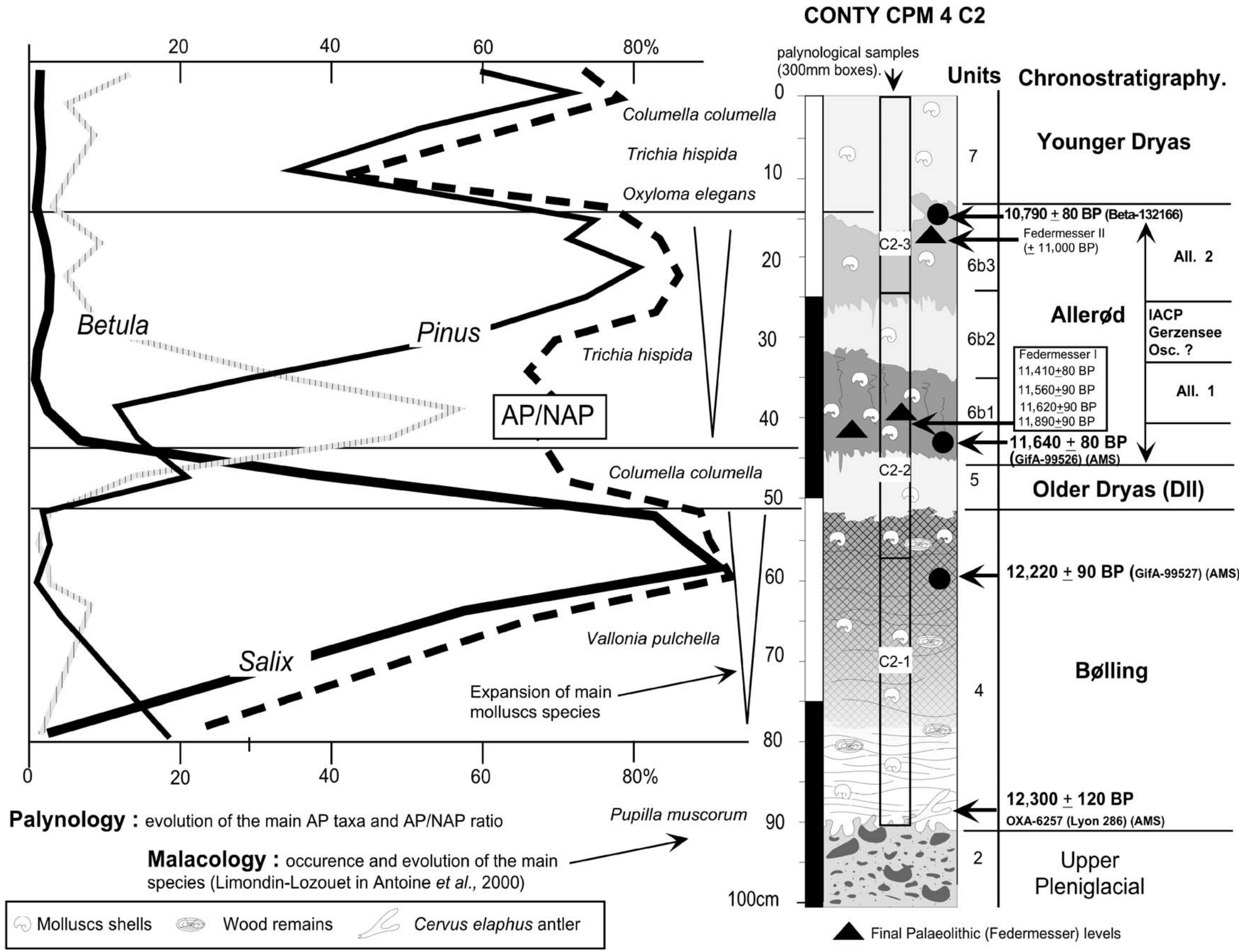



Fig. 3. Conty core CPM4–C2 (for location, see Fig. 2): stratigraphy, location of palynological samples, $^{14}C$ dating samples, and chronostratigraphical interpretation (for description of the units, see Fig. 2).

### 3.2. Allerød: transition to single meandering channel and deposition of silty organic material on the alluvial plain

In the Selle valley, the Allerød oscillation is distinguished by a net increase in the volume of alluvial sedimentation as well as a stronger reworking of slope bottoms. On the alluvial plain, sedimentation is characterized by the accumulation of organic calcareous silts on the margins of a wide single meandering channel. At Conty, this unit is subdivided into two organic horizons separated by a layer of lighter-coloured calcareous silt (Fig. 3, unit no. 6b2). The lower horizon yields the following dates: (AMS on charcoal) 11 420 $\pm$ 65 BP (Ly 285-OxA), 11 640 $\pm$ 80 BP (GifA-99526) and 11 410 $\pm$ 100 (GIFA-99525) (Figs. 2 and 3). On the basis of organic carbon contents (decreasing from 8% to 3%), and $^{14}C$ dates, the silty intercalation of unit 6b2 could be interpreted as the possible expression of the intra-Allerød Cold Period (Lehman and Keigwin, 1992) or the Gerzensee oscillation (Eicher, 1980, Lotter et al., 1992).

Laterally, in an abandoned channel (Channel IV), the upper part of the Allerød deposits grades laterally into a silty-peaty facies with plant remains (11,130 $\pm$ 80 BP and 11,080 $\pm$ 65 BP, Fig. 2). However, in the Conty sections, the threefold division of the Allerød unit is not clearly distinguishable from palynological or malacological evidence, despite an increase in *Pinus* and a decrease in *Betula*—leading to a crossing-over of the two curves—at the base of the median unit 6b2, along with a strong reduction in the total number of mollusc shells (Fig. 3). This pattern in the development of vegetation appears to be typical of the middle part of the Allerød, as seen in the palynological diagrams of the Netherlands (Hoek, 1997) and the Paris Basin (Munaut and Defgnée, 1997; Limondin-Lozouet et al., 2002). Nevertheless, it is likely that this intra-Allerød event was insufficiently long and cold to disrupt the vegetation that was already well developed during the first part of the Allerød. Broadly speaking, the Allerød is a period of relative filling of the valley by overbank deposits, with moderate elevation of the alluvial plain, and stability or weak aggradation of the channel bases (Fig. 5). The sedimentary budget is positive and can be estimated at approximately 300–350,000 $m^3/km^2$ in the Upper Selle valley.

At the same time, on the slopes (Conty, Saleux, and Belloy/Somme), this period is marked by the formation of a pedological horizon developed on loamy hillslope material including chalk granules (average thickness: 0.3–0.4 m). This "Allerød soil" is characterised by a brownish grey to greenish colour, hydromorphic features (iron oxide patches), a slight decalcification, low humus content (TOC: 0.9% to 1%), a dense network of root tracks as well as numerous earthworm biotubules and calcitic biospheroids (Canti, 1998), but no evidence of clay illuviation. The characteristics of this horizon are very similar to those of the "Allerød soils" in the chalk area of Kent, (Preece, 1995; Preece and Bridgland, 1999) and the Paris Basin (Pastre et al., 1997).

On the basis of numerous $^{14}C$ ages obtained from the Final Palaeolithic occupations which bracket this soil (Fig. 3), its formation took place within the Allerød oscillation between 11,800 and 11,000 BP (La Chaussée-Tirancourt-Prés du Mesnil, Étouvie and Conty) (Fagnart and Coudret, 1995, 2000; Fagnart, 1997b).

Nevertheless, according to the palynological data from Saleux, Hangest, Belloy/Somme, and the $^{14}C$ ages of the Final Palaeolithic (Federmesser) occupations discovered at several sites stratigraphically beneath the soil (11.8 to 11.6 $^{14}C$-kyr BP), this soil horizon was mainly formed during the *Betula* phase of the Allerød between 11,800 and 11,500 BP (Munaut and Defgnée, 1997; Munaut, 1998). Hence, this horizon is not developed on a previous deposit but is rather associated with a low rate of colluvial sedimentation ($<1$ mm/yr). As such, it can be considered as an upbuilding soil (Almond and Tonkin, 1999) recording the modification of vegetation during the climatic improvement of the early part of the Allerød. The generally accepted stability of the whole landscape during the Allerød (Pastre et al., 1997) must therefore be called into question, at least for the early stages in lower slope environments.

The deposition of the loamy colluvium at the base of the slopes during the Allerød could, in principle, indicate more marked humidity and/or a sparser vegetation cover at ground level. On the other hand, the abundant charcoals observed in the Allerød deposits of the Selle valley, as in many profiles of NW Europe (Van der Hammen, 1953; Van Vliet-Lanoë, 1987; Preece and Bridgland, 1999), suggest an increase in the frequency of forest fires (increased continentality), which could have played an important role in the surface disturbance of soils on the slopes. Lastly, the composition of the malacological assemblages coeval with Allerød also appears to reflect a relative drying up of environmental conditions in the bottom of the valley (Limondin-Lozouet and Antoine, 2001).

### 3.3. Younger Dryas: silty-calcareous filling and aggradation of the alluvial plain

During the Younger Dryas, the organic Allerød deposits are partially eroded (sediment blocks reworked by local cut-bank erosion at the base of the channel at Conty), then systematically covered by a thick filling of calcareous silts over the entire valley bottomland (1 m on average, Figs. 2 and 3, unit no. 7).

This filling is represented by "chalky mud" (detrital $CaCO_3$: 55–70%, TOC $<1$%), primarily formed by the action of freeze-thaw processes on chalky slopes.

Indeed, the experiments carried out by Lautridou (1985) and Lautridou et al. (1986) show that chalk blocks are totally broken down after 100–150 freeze-thaw cycles, producing a large amount of very fine chalk silt and mm-sized chalk granules.

These chalky sediments are then transported downslope during the spring melting of the snow cover, and laid down over the entire valley as overbank deposits (Vandenberghe et al., 1987). They are attributed to the Younger Dryas because of their sedimentological characteristics (abundance of detrital carbonate) as well as their malacological and palynological contents, which indicate a well-marked climatic deterioration situated between the end of the Allerød and the Preboreal (Limondin-Lozouet, 1997, 1998; Munaut, 1998; Antoine et al., 2000; Limondin-Lozouet and Antoine, 2001). This chronostratigraphic interpretation is corroborated by the date of 10 960 ± 85 BP obtained at the base of this unit at Conty (Fig. 2, channel IV), as well as by the age of the youngest Final Palaeolithic artefacts (Long Blade Technology), which often overly the surface of the calcareous silts (9800–10,000 $^{14}$C-yr BP, Fagnart, 1997b).

The geometry and the facies of these deposits indicate a regime of periodic overbank flooding in an alluvial plain with a single channel, leading to aggradation and progressive flattening of the valley bottom. The Younger Dryas channel facies is represented by silts and bedded freshwater coarse calcareous sands including layers of chalk granules and small flints. Although this facies is not recorded in the transect of Fig. 2 due to erosion by early Holocene incision, it is observed in a parallel profile located about 100 m away (Antoine, 1997a). Lastly, it is noteworthy that this important phase of sedimentation with high carbonate content characterizes the Younger Dryas throughout the whole Paris basin (Pastre et al., 1997, 2000) as well as in the chalk areas of South-eastern England (Preece and Bridgland, 1999). Moreover, we can observe that this kind of calcareous sediment (with very low organic carbon content) disappears definitively from the fluvial record at the end of the Younger Dryas cold phase, thus reinforcing their interpretation as cold period markers. In a general way, the climatic deterioration of the Younger Dryas results in a positive sedimentary budget in the Selle valley that can be estimated at approximately 1,000,000 m$^3$/km$^2$ (Fig. 5).

The Younger Dryas in the profiles of northern Europe is also represented by an abundant accumulation of fluvial overbank silts, then by aeolian dune sands in a strongly contrasted climatic regime (Vandenberghe and Bohncke, 1985; Munaut and Paulissen, 1973).

While the sediments of the Younger Dryas are present at Conty and Saleux in the Selle valley, they are partly eroded during the early Holocene in the Renancourt profile located farther downstream (Fig. 4, no. 8). This observation seems to indicate an enhancement of erosive processes downstream, as in the valley of the Somme.

### 3.4. Second phase of downcutting and early Holocene sedimentation

The transition with the Preboreal is above all characterized by a second major phase of downcutting in a large single-channel system with meanders (Fig. 2, Channel I, Fig. 4, channel III, and Fig. 5). This incision erodes off most of the calcareous silts of the Younger Dryas, especially in the lower reaches of the Selle River at Renancourt (Fig. 4) and at Saleux (Antoine, 1995), as well as in the middle Somme at Étouvie (Antoine, 1997a). It also moulds a new valley-bottom topography on which the last occupations of the Final Palaeolithic

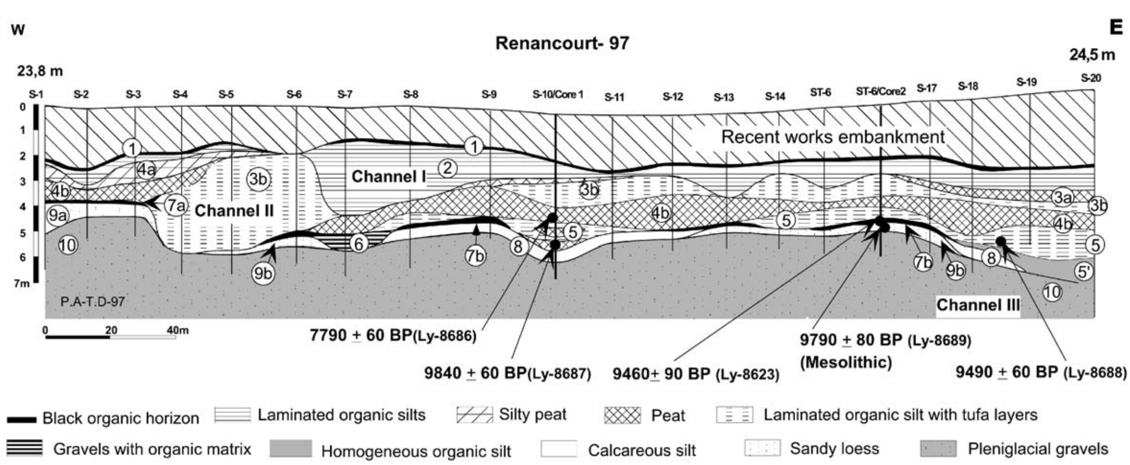
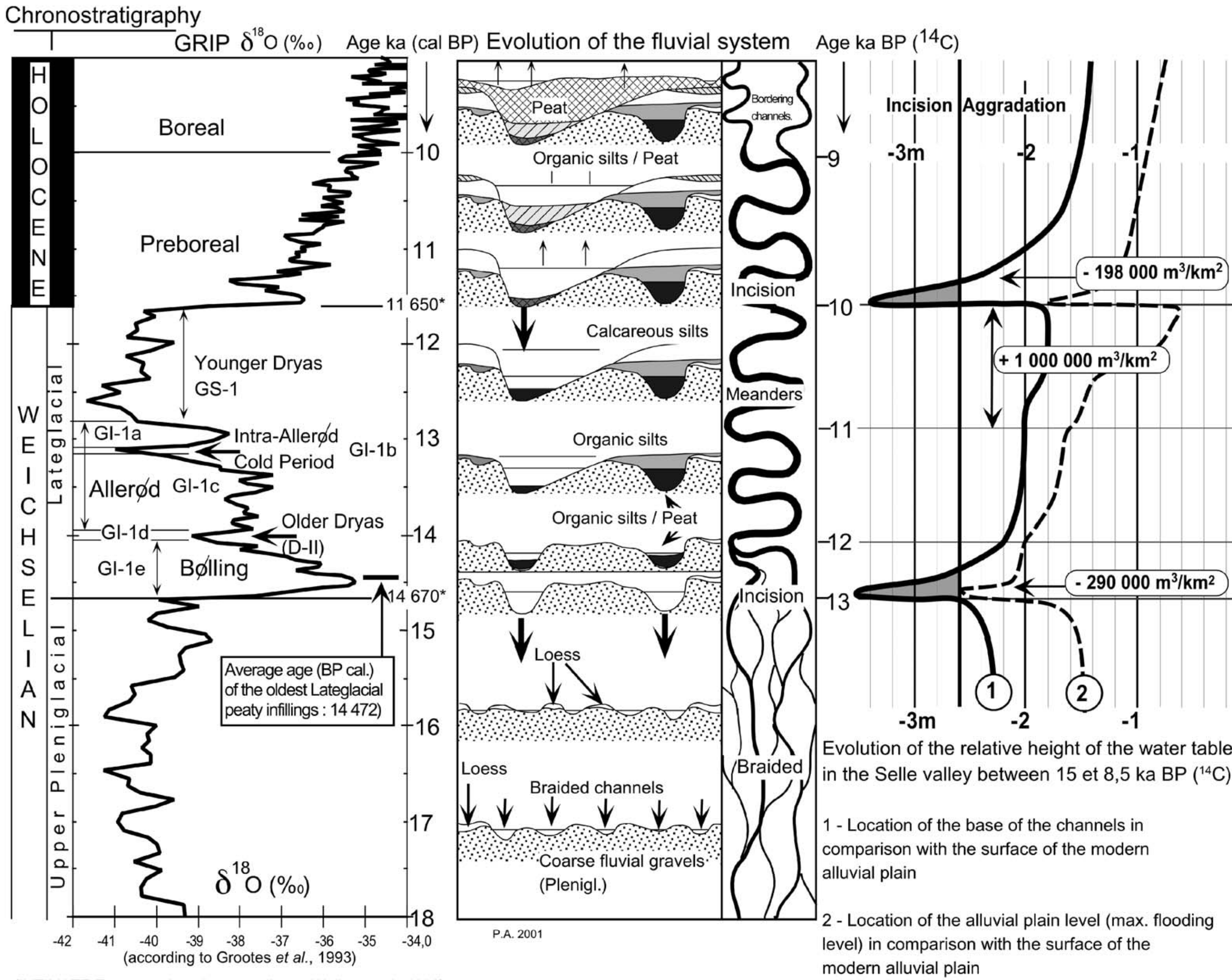



Fig. 4. Cross-section through the Selle valley at Renancourt (modified after Ducrocq & Antoine 1997; for location, see Fig. (1B)). (1) Organic marsh soil, (2) grey-green clayey silt, (3a) loamy peat, (3b) laminated organic loams with plant remains and tufa lenses, with lateral passage to peat in (4), (4a) loamy peat, (4b) typical peat, (5) organic loam with plant remains, tufa layers and peat lenses, (5′) homogeneous organic loam with plant remains, (6) flint gravels with organic silty matrix, (7a) organic loam with scattered chalk grains, (7b) homogeneous organic loam, (8) homogeneous calcareous silt, (9a) sandy loess with chalk grains and (9b) hydromorphic calcareous loam with chalk grains, with Mesolithic settlement at the top.




(INTIMATE Event stratigraphy according to Walker *et al.*, 1999).

* Ages Cal. BP of the beginning of the Bølling and of the Holocene according to Stuiver *et al.*, 1995 ; Taylor *et al.*, 1997).

Fig. 5. Modification of environments in the Selle valley during the Lateglacial, correlation with the $\delta^{18}O$ signal at GISP II, and preliminary assessment of sedimentary budgets for the Upper Selle area (modified after Antoine et al., 2000). (The sedimentary budgets are calculated on the base of the average thickness of the main units in the reference section at Conty extrapolated to a 1 km long area of valley. In addition, determination of the volume of the Younger Dryas deposits and of the eroded sediments at the beginning of the Preboreal is reinforced by data from the Saleux and Renancourt cross sections and by other intermediate observation points).

were established around 10,000 to 9800 $^{14}$C-yr BP (Fagnart, 1987, 1997a, b), in a still relatively open environment at the beginning of the Preboreal (large mammals dominated by *Equus caballus*, *Bos primegenius* and *Cervus elaphus*; Bridault, 1997). This important phase of downcutting, which occurred in most of the river systems of northern Europe, is related to the rapid onset of climatic improvement in the Preboreal (Vandenberghe et al., 1994; Huisink, 1998; Tebbens et al., 1999; Van Huissteden and Kasse, 2001). Prolonged landscape stabilisation during the Preboreal leads to the accumulation of the first clearly organic formations in the previously incised channel (Fig. 2, unit nos. 9 and 11), with the earliest peats dating back to 9900–9800 $^{14}$C-yr BP, as seen in the sequences at Étouvie (Reckinger and Munaut, 1995; Antoine et al., 1998; Munaut, 1998) and Renancourt (Fig. 4, unit no. 5). The first organic fluvial deposits of the Preboreal are characterized by a strong increase in the organic matter content of the sediment, but the silty fraction is still clearly dominant. Locally, in the Selle valley, the base of the Preboreal channel is coated by gravely deposits with an organic matrix derived from the erosion and reworking of the uppermost gravels on the valley floor (Fig. 2, unit no. 9).

## 4. Relations between climatic changes and fluvial processes

### 4.1. Downcutting at the beginning of the Bølling phase

The major modification in the fluvial regime of the Selle during the early part of the Bølling (before 12.4 $^{14}$C-kyr BP or $\pm$14,500 BP cal.) took place in an environmental and climatic context characterized by the following points:

#### 4.1.1. Local and regional data

*4.1.1.1. Modification of vegetation cover.* At the beginning of the Lateglacial, the vegetation cover is primarily dominated by *Betula* and *Salix* (more than half of the pollen grains, hygrophilous species excluded, Fig. 3). Nevertheless, it still contains shrubby plant taxa-dwarf forms in the case of some *Salix*-associated with *Artemisia* and Poaceae that are characteristic of steppe lands. Moreover, the peat accumulations are mainly made up of remains of *Salix* and *Betula* (Figs. 6 and 7). Compared to the conditions prevailing during the Upper Pleniglacial (loessic to grassy steppe), the early Lateglacial is characterized by the relatively poor development of arboreal vegetation in an environment still dominated by grasses (northern/open tundra). The malacological fauna of unit 4 (Figs. 2 and 3) reflect the progressive recolonization of the environment from 12.3 $^{14}$C-kyr BP onwards by a vegetation that becomes favourable for

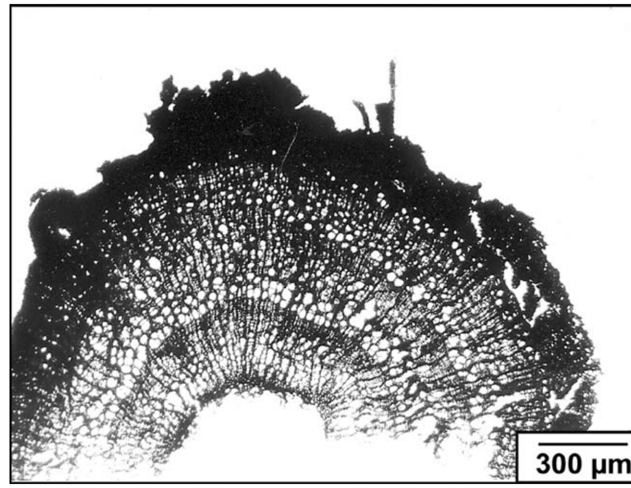



Fig. 6. *Salix* sp.: cross-section of a little branch (Photo J. & M. Dupéron).

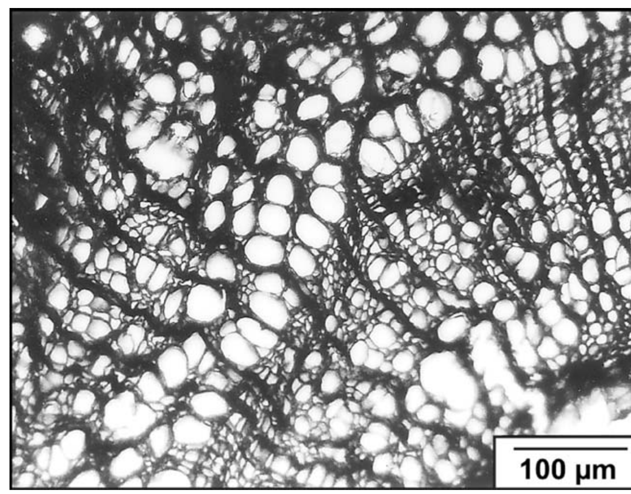



Fig. 7. *Salix* sp.: detail of the cross-section of Fig. 6 (Photo J. & M. Dupéron).

the development of malacofaunas typical of marshy biotopes (with *C. arenaria* and *Vallonia*). At the same time, changes take place in the large mammal fauna as found in the base of the channels at Conty, with the first appearance of *Cervus elaphus* (AMS on *Cervus antler*: 12 300$\pm$120 BP, OXA-6257-Lyon 286), indicating a markedly warmer climate. In addition, the study of beetle assemblages from the Bølling peat at Conty show that, during the infilling of the channels, the environment was very similar to the present day, with reconstructed maximum summer temperatures of about 15–16°C to 15–19°C (Limondin-Lozouet et al., 2002). The assemblages are dominated by marshy and aquatic species, but the presence of trees such as *Salix* and *Populus* is indicated by *Plagiodera versicolora*, *Melasoma populi*, *Phyllodecta tibialis* and *Chalcoides*. The very abundant macro-remains making up the Bølling "woody peat" include many large pieces (diameter up to

1.5 cm) of *Salix* and other big wood fragments (*Betula*/*Pinus* ?), showing that the arboreal component of the vegetation was composed of well developed shrubs and not low-growing arctic species such as *Salix herbacea*.

*4.1.1.2. Stabilization of slopes and cessation of aeolian sedimentation*. On the valley slopes, the development of the postglacial Bt horizon of Luvisol on loess (decarbonatization, leaching, and early stages of en-masse clayey illuviation) is mostly linked to the disappearance of permafrost (Van Vliet-Lanoë, 1987). This gradual development of the soils leads to a stabilization of the slopes, the limitation of lateral clastic supply and the confinement of discharge to the channels.

*4.1.1.3*. Disappearance of permafrost, and transition to a seasonal frost-type regime where the winter season gives rise to a thick snow cover (increase in precipitation and spring ice break-up related to melting of the snow cover). The disappearance of the permafrost has an important influence on the hydrological and sedimentary budget: it allows the re-establishment of vertical drainage in the soil and leads to a cessation of gelifluction on the slopes. This latter phenomenon caused the clogging up of the valley by lateral coarse sediment supply during the late Pleniglacial (Hoek, 1997; Huisink, 1998).

*4.1.2. Global context*

*4.1.2.1. The beginning of the Lateglacial is marked by:.*

(a) A sharp increase in temperatures, according to GRIP and GISP II $\delta^{18}$O data (Dansgaard et al., 1993; Grootes et al., 1993; Stuiver et al., 1995, Fig. 5), as well as coleopteran assemblages from NW Europe (Lowe et al., 1994; Walker et al., 1994).
(b) A modification of oceanic and atmospheric circulation in the North Atlantic, and a rapid shift of the polar front linked to reactivation of the return of warm surface water towards the north (Cortijo et al., 2000).
(c) A strong increase in precipitation regime, occurring within about a decade according to snow accumulation rates at GISP2 (Alley et al., 1993). This may have occurred over a longer time-span ($\pm 1$ ka), as indicated by quantitative climate reconstructions from pollen (Pons et al., 1992; Guiot and Couteaux, 1992) or organic carbon $\delta^{13}$C (Hatté et al., 2001). On the other hand, following this initial intense precipitation event, the second part of the Bølling appears to become increasingly drier (Bohncke and Vandenberghe, 1991; Walker et al., 1994, Magny, 1995).

A sharp increase in temperatures at the beginning of the Lateglacial, particularly during the summer, is thus clearly recognized at Conty from the coleoptera of the Bølling peat (unit no. 4). This result is comparable with temperatures in the Somme basin at the present day, and with values obtained from the Bølling sediments of the Oise River in the Paris Basin (Limondin-Lozouet et al., 2002), and from Holywell Coombe (Preece and Bridgland, 1999).

Generally speaking, fluvial incision and transformations of fluvial morphology are conditioned by diminishing sedimentary load and increasing discharge of the watercourses (Schumm, 1977; Vandenberghe, 1995; Collinson, 1998). The synthesis of data for the beginning of the Lateglacial in the Selle valley shows that a change takes place in the balance between sedimentary load and transport capacity in an environment characterized by the cessation of aeolian sedimentation, the disappearance of permafrost, and initial soil formation at around 16–15 kyr (cal. BP). This is associated with modifications in the vegetation components towards the end of the Pleniglacial.

On the other hand, according to the palynological data obtained at Conty (Munaut in Antoine et al., 2000), the onset of downcutting definitely pre-dates any marked extension of the vegetation. Indeed, the expansion of *Salix* and *Betula* began only within the upper two thirds of the peat interval of unit 4 (Fig. 3). This evolution is moreover correlated with a progressive change in the sedimentation through unit 4 from calcareous organic laminated silts in the lower part (TOC: 2%) to typical brown woody peat at the top (TOC: 15%).

Taking in account these data and $^{14}$C dates from the lower part of unit 4 (Fig. 2), modifications in the vegetation cannot alone represent the triggering factor responsible for downcutting. Nevertheless, regarding the impact of vegetation on landscape stability, it is clear that a continuous grassy environment (shrub tundra cover) can be very effective in protecting soils from erosion and that well developed forest is not required to play this role.

In addition, a comparison with global data shows that this major landscape modification (incision) at the onset of the Lateglacial is coeval with the earliest period of rapid temperature rise indicated by GRIP $\delta^{18}$O data (base of GI-1e, Walker et al., 1999, see Fig. 5). This very rapid temperature rise (within less than 100 yr) is contemporary with a change of climatic regime associated with a sharp increase in precipitation over Western Europe (Guiot and Couteaux, 1992; Cortijo et al., 2000).

Thus, we can interpret the triggering of a rapid phase of downcutting in terms of an increase in the effective water discharge—and spring flooding in particular—in a still seasonal climatic regime (importance of snow cover and spring melting, linked to an increase in seasonality). In parallel, the melting of the permafrost ice during the sudden climatic improvement at about 13.0 $^{14}$C-kyr BP produced a huge amount of water that is likely to have enhanced the water discharge and downcutting of the river. This scenario is compatible with the pedological data (decarbonatization/decalcification, leaching of loess, vertical drainage), which imply a cessation of hillwash dynamics and solifluction, as well as an increase in the infiltration and drainage of the soils. However, such processes can only be sustained in an environment where the sedimentary inputs are progressively reduced by soil development that has likely began prior to the change in the vegetation components.

The development of shrub then of forest cover then amplifies the soil stability during the second half of the Bølling and into the Allerød by stabilizing the slopes and channel banks. Nevertheless, compared to the late Bølling period, when sedimentation is limited to the channels (around 12.2–12.3 $^{14}$C-kyr BP), the Allerød seems to be a period of relative instability of the environments and, more particularly of the slope bottoms (colluvial deposition and upbuilding soils). In addition, according to the cross-section at Conty, there is an increase in sedimentary budget in the valley between 11.6 and 11 $^{14}$C-kyr BP (more marked flood dynamics) (Fig. 5).

Finally, the sedimentary shift attributed to the Older Dryas (Dryas II) is marked by an episode of alluvial accumulation (overbank deposits), with predominant detrital carbonates, taking place in an environment with relatively little destabilization of the vegetation. The triggering factor of this rapid response to a very short climatic event is most likely the return of strong winter freezing, which involves a much more intense frost-shattering of chalk on the slopes. Lastly, it is noteworthy that the Older Dryas is represented at Conty by the same type of calcareous silt as observed in the case of the Younger Dryas.

### 4.2. Major valley filling during the Younger Dryas

During the Younger Dryas, the Selle valley was the site of thick calcareous silt (chalk mud) sedimentation primarily derived from chalky slopes in a cold and wet climate (especially in the early part of the Dryas, Limondin-Lozouet and Antoine, 2001). These chalk muds are formed by frost-shattering on the limestone slopes (see Section 3.3), then being concentrated in the valley bottom and mixed with fluvial silt (about 60–70% $CaCO_3$/30–40% silt) during the melting of the snow cover in spring and early summer. They are subjected to intense bioturbation related to the grassy vegetation, which progressively colonizes the sediments as soon as they are deposited (destruction of sedimentary structures). This interpretation is compatible with the results of pollen analysis studies in the Netherlands, which show the strongly contrasted character of the climate at the beginning of the Younger Dryas (cold and wet), with average summer temperatures estimated at 10–11°C (Vandenberghe et al., 1987; Kasse et al., 1995). In the Selle valley, the geometry and facies of the Younger Dryas calcareous overbank deposits point to a discharge regime of periodic flooding from a single channel onto the alluvial plain. At Conty, the Younger Dryas channel filling is made up of calcareous silts and coarse sands containing *Trichia hispida* but also high number of the aquatic species *Ancylus fluviatilis*. It is represented by clearly laminated sediments with organic intercalations at its base (Antoine, 1997b).

Broadly speaking, the accumulation of the calcareous overbank deposits involves a rising and progressive planation of the surface of the alluvial plain (planar surface at the end of the Younger Dryas, Fig. 2, no. 7). The decline of the vegetation cover is expressed mainly by a relative decrease in the abundance of *Pinus* and *Betula* pollen and by a dramatic impoverishment of malacofaunas. In such a climate characterised by a generalized fall in temperatures, which may locally involve the return of periglacial conditions in the Netherlands (cryoturbation, cracks, Hoek, 1997; Isarin, 1997; Huijzer and Isarin, 1997) and locally in the Paris Basin (Pastre et al., 1997), this strong increase in the abundance of sedimentary carbonates in the valley may be primarily due to the impact of freezing on the nearby chalky slopes. Slope erosive processes are therefore located preferentially in the immediate vicinity of the valleys, on the abrupt chalky slopes where frost-shattering is most effective.

### 4.3. Downcutting at the beginning of the preboreal

In the Selle valley, as in the Somme, a second major phase of downcutting is recognized at the transition between the Younger Dryas and the early stages of the Preboreal. In fact, the base of the silty organic and then peaty material filling the Holocene channel is dated at 9900 $^{14}$C-yr BP at Étouvie (Antoine et al., 1998) and 9800 $^{14}$C-yr BP at Renancourt (Ducrocq and Antoine, 1996) (Fig. 4). This second phase of downcutting occurs during a period of climatic amelioration, inducing recolonization by trees (in particular *Betula* and *Pinus*) following the Younger Dryas Stadial.

From a global climatic point of view, this transition is contemporary with a sudden increase in temperature according to the GRIP and GISP2 $\delta^{18}$O records (Grootes et al., 1993; Stuiver et al., 1995; Taylor et al., 1997). It is also reflected in the quantitative climate reconstructions from pollen and coleoptera (Ponel and Coope, 1990; Guiot et al., 1992; Walker et al., 1994), molluscs (Rousseau et al., 1998), as well as $\delta^{18}$O data from lake sediments in Europe (von Grafenstein et al., 1999).

In addition, the rearrangement of oceanic and atmospheric circulation in the Atlantic area is similar to that at the beginning of Bølling, with an especially rapid rise in snow accumulation rate over Greenland (Alley et al., 1993), and an increase in precipitation in continental records (Guiot et al., 1989; Guiot and Couteaux, 1992; Pons et al., 1992). Nevertheless, as for the Bølling, the initial precipitation peak at the boundary between Younger Dryas and Preboreal is followed by globally drier conditions during the Preboreal (Magny and Ruffaldy, 1994).

Compared with the Upper Pleniglacial/Bølling transition, where downcutting occurs in parallel with an important modification in river morphology, renewed downcutting at the beginning of the Holocene takes place in a single channel with large meanders that existed already during the Younger Dryas (channel facies description: Antoine, 1997b). In cross-sections at Conty, Saleux, Renancourt and Étouvie, we observe a clear deepening of the channel (from 1 to 3 m), with a varying degree of lateral migration that, in most cases, causes the erosion and reworking of the Younger Dryas channels facies and variable proportions of the overbank deposits. This explains the very poor preservation of Younger Dryas channel facies in the lower Selle valley and more generally in the Somme (Antoine, 1997b).

At the beginning of the Preboreal, downcutting can be interpreted as the result of a sudden enhancement in sediment transport capacity linked to global climate change (shift to more oceanic climate, increase in temperature and precipitation). This induced a sharp increase in run-off and transport capacity, combined with a reduction in the sediment supply. This phase is correlated with the development of the largest channel width ever attained, during Lateglacial and Holocene times, which is locally up to five times larger than the present-day channel of the Selle (15–20 m max. as against 100 m for the Preboreal Channel: Fig. 2, Channel I).

The reduction in the sediment supply may be related to the following factors:

(1) A cessation of slope erosion linked to freeze-thaw processes is shown by the definitive disappearance of chalky-slope sediment input into the river system.
(2) Rapid expansion of arboreal vegetation characterised by two successive phases: a Betula phase from $\pm 10$ to 9.5 $^{14}$C-kyr BP, then a *Pinus* phase from 9.5 to $\pm 9$ $^{14}$C-kyr BP (Munaut, 1998).
(3) Deepening of the soil profile by leaching processes (brown leached soil/Luvisol on the loessic cover).

According to $^{14}$C dates from the basal part of the organic infillings in the Holocene channel, incision was completed before the *Pinus* phase in the Selle and in the Middle Somme, and took place during the first part of the *Betula* phase, as observed in other rivers of the Paris Basin (Pastre et al., 1997). Thus, we can assume that a time lag of some centuries occurred between the very sharp rise in temperature and precipitation recorded in the global climatic data at $\pm 10$ to 10.2 $^{14}$C-kyr BP (11.55–11.65 kyr BP cal.) and the response of the vegetation. The response of the river system is marked by faster incision occurring in an environment where AP percentages remain relatively low. Finally, if the development of arboreal vegetation is not the triggering factor, valley incision could result from a strong increase in run-off over a short period of time (1 to 2 centuries), combined with a sudden halt in chalky sediment supply (cessation of freeze-thaw dynamics at the end of the Younger Dryas).

On the other hand, during the second part of the Preboreal (9,5–9,0 $^{14}$C-kyr BP), there is evidence for a sharp increase in evapotranspiration linked to the spread of arboreal vegetation, especially of *Pinus* (Munaut, 1998), and of a (global) drier climate (low lake levels: Guiot and Couteaux, 1992; Magny and Ruffaldy, 1994). In parallel with this, some pedological processes partly ceased during the Younger Dryas (impact of seasonal freezing) and started again later leading to deepening of the soil profile. This stabilization of slopes during the Preboreal and increased dissolution of the chalky substratum is also clearly demonstrated by the first appearance of calcareous tufa at the base of the Holocene channels (Étouvie, Conty, Fig. 2 no. 10). This phenomenon can also be observed in many river valleys of the Paris basin, for example in the Marne and Oise (Pastre et al., 1997), and is consistent with the age of the first tufa at Holywell Coombe (9760 $\pm$ 100 $^{14}$C-yr BP, Preece and Bridgland, 1998).

The development of the vegetation only becomes effective at a later stage of the Preboreal. On the other hand, above a certain threshold, evapotranspiration clearly limits the transport capacity of the river and prevents the process of downcutting. When the vegetation cover reaches a more advanced stage of development (Preboreal and Boreal), the increase in evapotranspiration and greater stabilization of the soils minimizes the processes of runoff and supply of fine detrital material into the valley. This tendency is clearly illustrated by the rapid transition often observed in the Preboreal channel fillings between increasingly organic silts and typical peats (Conty, Saleux, and Renancourt).

This evolution causes a rapid filling by organic silts and peat, and the disappearance of the large Preboreal channels. Later, during the Boreal and early Atlantic, this process is enhanced leading to a generalized filling up of the valley by peat marsh formations between about 8000 and 7000 yr BP. In this new valley configuration, the groundwater circulates primarily in the coarse alluvial aquifer (Pleniglacial gravels) and within the peat marsh. The surface water flow is apparently very strongly reduced, with only small localized shallow lateral channels remaining at the junction between the base of the slope and the peat marsh (Antoine, 1997a) (Fig. 5).

## 5. Conclusions

In this study, we present results on the modifications of sedimentation and river morphology in the Selle valley. Taking this into consideration, along with data on climatic variations at the scale of Western Europe and the North Atlantic, we can propose the following conclusions:

1. The major phases of downcutting, which lead to the remoulding of valley morphology, occur systematically at the very onset of episodes of rapid climatic warming that characterize the beginning of the Lateglacial and the Holocene ($<100$ yr). In addition, the modifications observed in the Selle valley and more widely in the Somme Basin are contemporary with those described in other river systems of Western Europe, although they are generally more extensive.
2. Major downcutting is apparently characteristic of short periods of very rapid climatic change from cold and arid conditions towards a much more temperate and humid climate with a still strongly seasonal discharge regime. During these short episodes, the precipitation is much greater in amount and shows an irregular distribution pattern.
3. The climate change at the beginning of the Lateglacial at about 13 $^{14}$C-kyr BP leads to downcutting because of the sudden increase in water discharge contemporaneous with a sharp reduction in slope sediment supply. This reduction of sediment supply results from a cessation of aeolian sedimentation, the disappearance of permafrost and the development of soils linked to the sudden change in climatic conditions occurring at that time. In parallel, a sharp increase in precipitation and (or) some modification of the precipitation regime (increased seasonality) could enhance the carrying power of the rivers and lead to incision.
4. Even if variations in vegetation density can have an impact on the stability of soils and thus lead to a reduction of clastic supply from the catchment area, according to the data from the Upper Selle River, this factor does not seem to trigger downcutting. In particular, at the beginning of the Lateglacial, the downcutting clearly pre-dates the spread of the arboreal (shrub) vegetation cover that is recorded only at about 12.3 to 12.2 $^{14}$C-kyr BP within the peat infilling of the newly created channels. Although this unconventional interpretation is clearly supported by the $^{14}$C and palynological data from the Selle, it needs to be validated by investigating profiles in other small river valleys of NW France.
5. The climatic improvement and rate of downcutting in the earliest Holocene are comparable to the Bølling, but the impact of vegetation at this stage is likely to be much more important (cf. widespread tree cover). Nevertheless the new spread of the vegetation cover does not really take place until the middle Preboreal at around 9.5 $^{14}$C-kyr BP. This post-dates the main incision phase that is older than 9.9 to 9.8 $^{14}$C-kyr BP. Compared with the situation at the beginning of the Lateglacial, the incision at the start of the Holocene takes place during a shorter time span within a pre-existing large single meandering channel.
6. Then, during the younger part of the Preboreal, the drier climate and the expansion of tree cover causes an important increase in evapotranspiration, which rapidly limits water flow towards the valley and leads to a decrease in the clastic (loamy) sediment supply as well as the rapid development of peat during the early Holocene.
7. Relatively long cold phases such as the Younger Dryas (lasting $\pm 1000$ yr) are characterized by a distinct local disruption of the landscape and a particularly well-marked impact on the chalky slopes (role of frost shattering). This situation also arises during much shorter phases, such as the Older Dryas. In such cases, on the contrary, the local vegetation in the valley does not undergo any marked change, probably because its response time is too slow. In small river valleys on chalk bedrock, the occurrence of highly calcareous overbank deposits within the fluvial record may be considered as a signature of cold phases during the Lateglacial. On the other hand, the response of the Selle and Somme rivers to cold climatic conditions during the Younger Dryas never leads to the return of braided river systems as seen in some larger rivers such as the Meuse, where internal factors including longitudinal gradient and sediment grain-size play a more important role.

Finally, the upstream courses of small river valleys in north-western France offer very favourable environments for studying the impact of Lateglacial climatic variations in a continental setting. However, in comparison with lake sediments, the periods of non-deposition and erosion in fluvial environments make it much more difficult to obtain uninterrupted records (e.g. from a core sample). On the other hand, the discontinuous nature of such environments is a major advantage since it enables a clear identification of the sedimentological and morphological changes, as well as a hierarchical treatment of the different recorded climatic crises.

## Acknowledgements

In parallel to the work undertaken as part of preventive archaeology operations in collaboration with the staff of AFAN, this research on the evolution of the river systems of the Somme Basin during the Lateglacial and Holocene was supported by the CNRS within the framework of the Palaeoenvironments and Hominids programme (1997–2000), and since 1995 by the General Council of the Somme Department (Valley Bottoms programme). Dr. M.S.N. Carpenter translated the manuscript. The author thanks L. Tebbens and one anonymous referee for their constructive review of this paper.

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