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Modelling the long-term fluvial erosion of the River Somme during the last million years

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ABSTRACT

A process-based model that simulates fluvial erosion in the River Somme Valley over the last million years is presented here. The model takes into account lithology and climatic influences and allows the simulating of undercapacity and overcapacity sediment transport behaviour. The model has been calibrated to a family of terraces within the River Somme Valley. When matched to this field data, simulation trials suggest that bedrock incision occurred principally from 120 to 60–40 kyr during the last climatic cycle and before the last glaciation. The impact of a progressive tectonic uplift (*c.* 60 m over *c.* 1 million years) on the River Somme has also been studied here. Extended over a longer period of time, the simulations suggest that 1 million years ago the profile of the River Somme had a lower slope gradient than today, with little relief throughout the Paris Basin.

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Introduction

Fluvial processes are controlled by numerous parameters. In long-term erosion, the main forcing factors are tectonic, lithological, climatic and sea-level variations (Schumm, 1977; Veldkamp and Tebbens, 2001). As in other non-linear complex systems, an apparent gradual change of an external variable may trigger a series of responses in fluvial systems due to self-organization of the system (internal controls) (Veldkamp and van Djike, 2000; Bogaart *et al*., 2003). In the last decade, the long-term fluvial evolution and associated climatic changes have been studied in different geographical areas. There are numerous archives of palaeorivers in North-West Europe for the last million years (Antoine, 1994; Vandenberghe *et al*., 1994; Bridgland, 2002). Palaeohydrological studies indicate that fluvial systems react strongly to climate change due to associated complex response dynamics in their drainage basin (Veldkamp and van Djike, 2000).

Despite the information from fluvial records being available, there are important gaps in data concerning long-term (*c.* 100 ka–1 Myr) fluvial evolution. The discontinuity of the fluvial system archives is due to lack of data following erosion and non-deposition. Fluvial erosion has been documented in different periods during the last climatic cycle in NW Europe (Antoine, 1994; Vandenberghe *et al*., 1994; Antoine *et al*., 1994, 2000; Pastre *et al*., 2000; Van Huissteden *et al*., 2001; Fig. 1). Nevertheless, the fluvial evolution is not precisely understood because of the superimposition of internal and external controls and also of the difficulty to obtain precise river deposits dating. How can we understand long-term fluvial river evolution when only discontinuous field data are available? How can we quantify tectonic uplift when this uplift is low and no structural evidence of ground deformation actually exists?

The 'biogeoprospective project' initiated by ANDRA (National Radioactive Waste Management Agency) aims at providing a better understanding of the past environmental change, to predict environmental evolution. Such a prediction should be made by bringing together climate and geomorphological modellers and specialists from various palaeo-ecological disciplines. This study is part of this project, focusing on long-term river erosion modelling. The aim of this paper is to enhance our comprehension of the processes controlling the fluvial response to climatic changes; it is also to understand the struggle between forcing factors with precise time control. To facilitate comparison between internal and external controls, we will develop a fluvial system model, based on simplified climatic, tectonic and sea-level inputs. To gain quantitative insight into the validity of the simulated system dynamics, we have calibrated the model on the Somme settings.

The River Somme

The River Somme flows in a small valley situated in the north of France, characterized by a drainage basin of *c.* 5800 km^2 and has developed on an homogeneous chalk bedrock of upper Cretaceous age (Fig. 2) (Mégnien, 1980; Antoine, 1994). From the present coastline to the Greenwich deep, the river incised Tertiary sands and clays. Downstream from the Greenwich deep, the river flowed over Jurassic limestone when the Channel emerged at a time of low sea level (Lericolais, 1997; Antoine *et al*., 2000).

Within the Paris Basin, the quantification of the uplift rates is often based on the fluvial terraces, similar to the Maas or Thames Valleys. The well-documented terrace system of the River Somme (Fig. 3) (Commont, 1910; Bourdier, 1969; Sommé *et al*., 1984; Antoine, 1994) and the dating of these fluvial records (Fig. 4) (Bates, 1993; Laurent *et al*., 1994; Antoine

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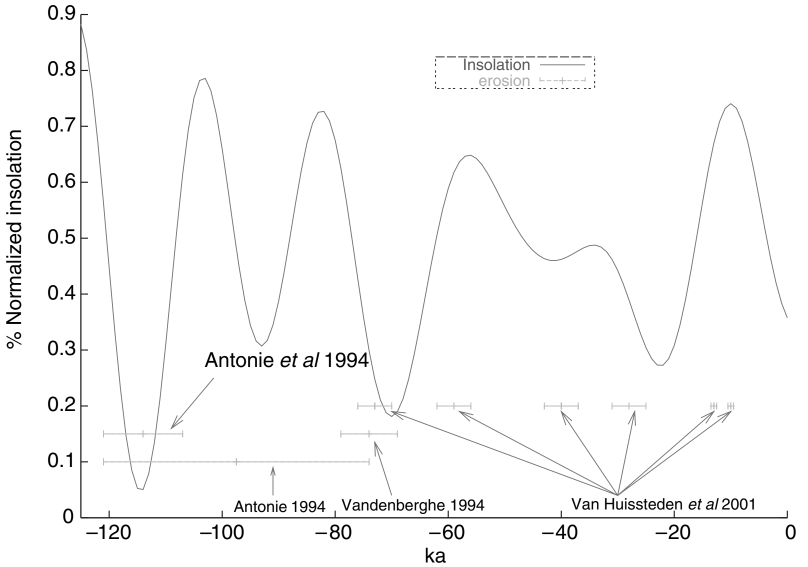

Fig. 1 Synthesis of erosion phases in NW Europe during the last climatic cycle (Antoine *et al*., 1994; Antoine, 1994; Vandenberghe *et al*., 1994; Van Huissteden *et al*., 2001). Comparison with a normalized insolation curve calculated at 65°N (from Berger and Loutre, 1991).

et al., 2000) offer a good opportunity to compare the model with field data. It constitutes a relatively well-preserved record of fluvial evolution over the last million years. The fluvial terraces of the Somme have been identified only in the downstream part, between the present coastline and Amiens. For this reason, it is not easy to offer a complete geometric profile for the palaeo-River Somme.

The Somme fluvial terrace geometry was interpreted by Antoine (1994), as the result of an homogeneous uplift of the valley. The homogeneity of the uplift not only in time, but also in space, has been contested (Van Vliet Lanoe *et al*., 2000). Nevertheless, the 55–60 m Myr^{-1} uplift rate proposed by Antoine *et al.* (2000) for the Somme Valley is of the same order as the 70–100 m Myr^{-1} non-homogeneous uplift rate of the Thames Valley (Maddy *et al*., 2001) and the 70 m Myr^{-1} uplift rate of the Maas valley in the Ardennes (Tebbens *et al*., 2000). Uplift is often considered an important ingredient in terrace formation and is a well-documented phenomenon (Jones *et al*., 1999; Schumm *et al*., 2000).

The tectonic deformation of the Paris Basin during the last few million years is not highly significant and is generally expressed as a uniform uplift. Some faults are parallel to the present coastline in this area (Antoine *et al*., 2000) and to the northern French coastal cliffs. However, there is also a currently active fault parallel to the Somme Valley (Goffé *et al*., 1998). Quaternary marine terraces have been observed in Sangatte (Balescu *et al*., 1992) and also in other parts of the current French northern coastline (Antoine *et al*., 1998, 2000).

Processes and field data

During the last climatic cycle, many periods of erosion appear to have taken place in NW Europe, based on field evidence (Fig. 1) (Pastre *et al*., 2000; Van Huissteden *et al*., 2001). The last 20 ka were thoroughly studied (Pastre *et al*., 2000), however, it is more difficult to obtain accurate data for the previous periods (120–20 ka). During Marine Isotope Stage 3, fluvial evolution did not generally produce a clear response to climatic change. During Stage 3, phases of aggradation occurred, however, the period of maximum aggradation varied between sites in Northern Europe (Van Huissteden *et al*., 2001). In several sites, the earliest Marine Isotope Stage 3 deposits are also lacking and an incision phase of Hengelo interstadial age was found in the Netherlands (Van Huissteden *et al*., 2001). Marine Isotope Stage 4 was mainly a time of intensified fluvial incision in Northern Europe. This incision is more pronounced in NW Europe than in NE Europe (Van Huissteden *et al*., 2001). The precise evolution at a time-scale of 120–20 ka is unclear because of the lack of information. Nevertheless, the main periods of fluvial incision of the bedrock of the Somme probably happened during the early glacial phases (Antoine, 1994) (Fig. 1).

The long-term climatic control of fluvial evolution is one of the most evident, however, it is also one of the more difficult long-term processes to model, because of the lack of data used to quantify fluvial dynamic parameters (e.g. water discharge or sediment supply). Not all long-term river modelling studies take this point into account. Despite uncertainties, some information can be extracted from physical processes. The most important impact of long-term climatic variations on fluvial systems is the evolution of water and sediment flux over time (Veldkamp and Tebbens, 2001). Previous studies considered that water discharge was greater during interglaciation periods than at other times because the rainfall rate was higher (Veldkamp and Tebbens, 2001). Nevertheless, to quantify the water discharge, it is necessary to take into account the rainfall as well as the evapotranspiration and infiltration. It is therefore necessary to more precisely analyse the evolution of water discharge over time and to quantify the consequences of changes in water and sediment flux.

The water flux Q depends on rainfall R, infiltration I, evapotranspiration EVT and snow storage N_{snow}:

$$Q = R - I - \mathrm{EVT} - N_{\mathrm{snow}}. \quad (1)$$

If rainfall is more significant during warmer periods than during colder ones (Guiot *et al*., 1989), infiltration and evapotranspiration are also more significant. During interglacials or relatively warm periods, the subsoil is highly permeable and most runoff will occur subsurface. During glacial or relatively cold periods, the subsoil is frozen for part of the year and is no longer capable of massive subsurface flow. Most run-off will occur as overland flow and in more intense events.

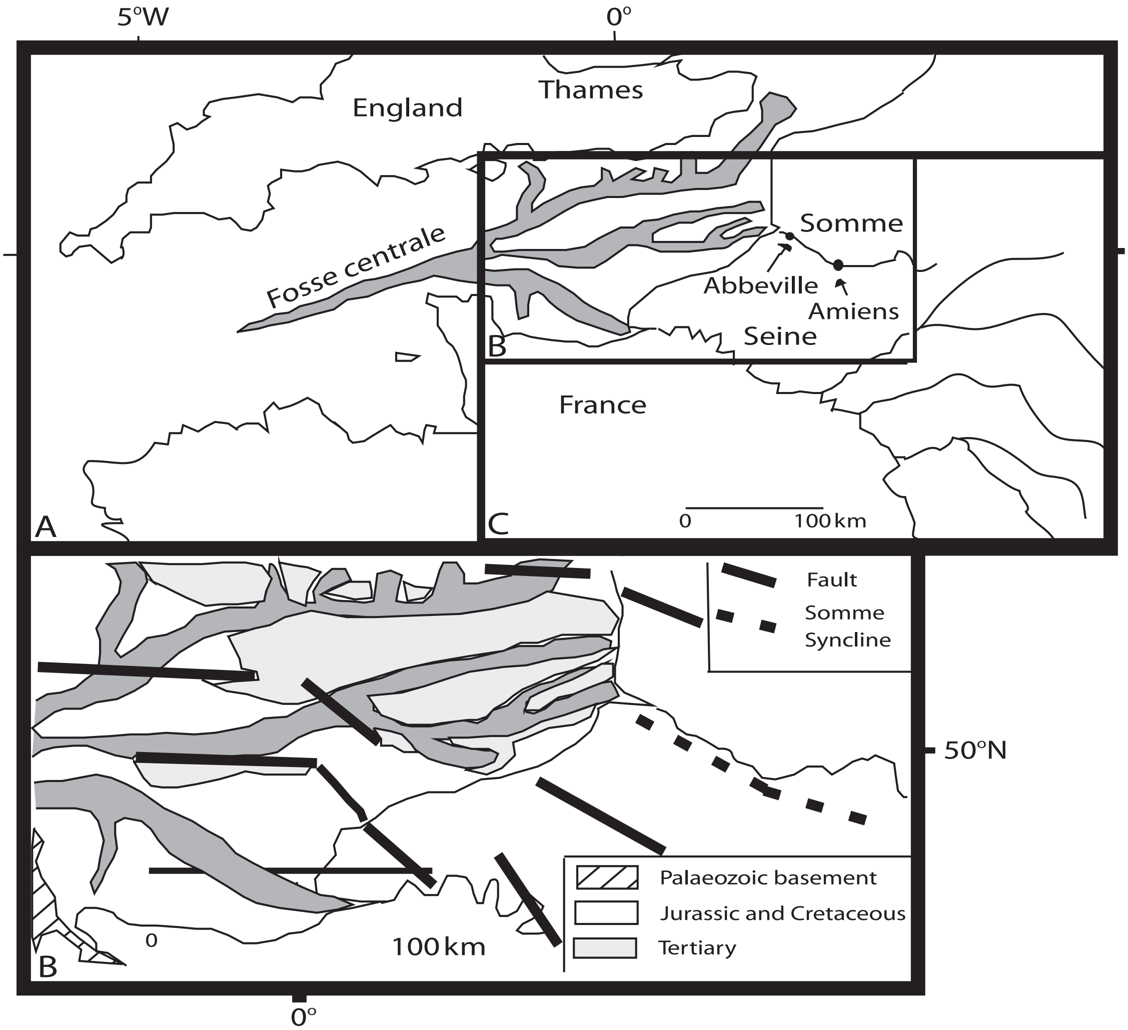

Fig. 2 Location map of the River Somme (from Antoine *et al.*, 2000). The Paris Basin is in the area C.

It is then necessary to more precisely quantify infiltration and evapotranspiration, to simulate the variations of the river flow during a climatic cycle.

Evapotranspiration can be calculated for NW Europe as a function of rainfall R (mm yr^{-1}) and mean annual temperatures T (°C) by the equation proposed by Bogaart and van Balen (2000):

$$\mathrm{EVT} = \frac{R}{\sqrt{0.9 + (R/L)^2}} \qquad (2)$$

with

$$L = 300 + 25T + 0.05T^2.$$

During an interglacial period in NW Europe, the rainfall was around 1000 mm yr^{-1} and the mean annual temperature is estimated between 10 and 15 °C (Guiot *et al.*, 1989; Fauquette *et al.*, 1999). Under these conditions, the EVT is around 490–580 mm yr^{-1}. Considering that the infiltration in the Paris Basin during interglacial periods was equivalent to the present infiltration rate of 200 mm yr^{-1} for a rainfall of 1000 mm yr^{-1}, the value of the water flux Q was of 200–300 mm yr^{-1}.

During the last glacial period in NW Europe, the mean rainfall was around 500 mm yr^{-1} (Guiot *et al.*, 1989; Fauquette *et al.*, 1999). If we consider during this time a mean annual temperature of 5 °C, the calculated EVT is of 330 mm yr^{-1}. Considering that infiltration was negligible during this period, we obtain a possible water flux Q of 170 mm yr^{-1}. This value is of the same order as the water flux calculated for warmer periods.

Nevertheless, fluvial processes are not controlled by mean annual flux but by ‘characteristic’ water flux, depending on the maximum water flux (Pomerol and Renard, 1997; Gargani, 2004a). Peak discharges that significantly modify the morphology of the valley are not proportional to rainfall for the last climatic cycle for NW Europe. During glacial periods in winter, the climatic conditions were favourable to snow storage and water flux were probably very small. The water flux then increased during glacial periods in spring or summer, due to the snow melting. On the other hand, during NW Europe interglacial periods, the water flux was distributed in a more homogeneous way throughout

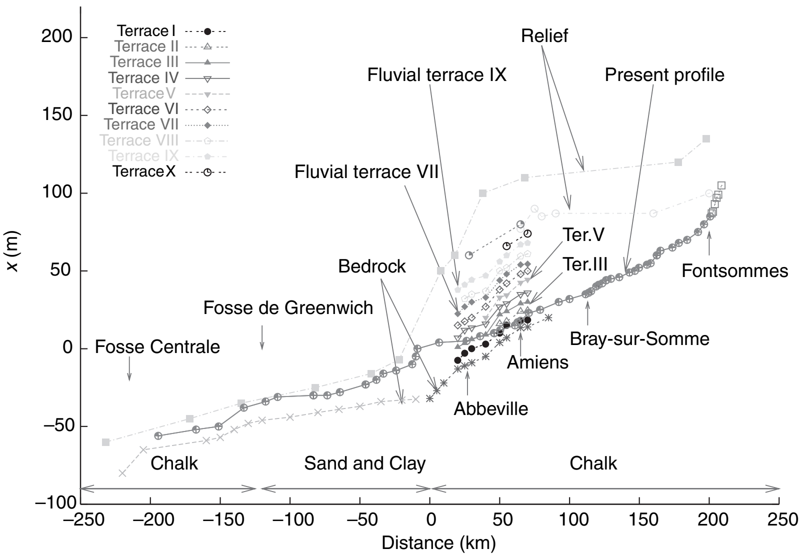

Fig. 3 Main geometrical feature of the River Somme: profile, terraces between Abbeville and Amiens, bedrock, relief and lithological changes (Alduc *et al.*, 1979; Antoine *et al.*, 2000). The coast is at the distance 0 km. Age of fluvial terraces are given in Fig. 4.

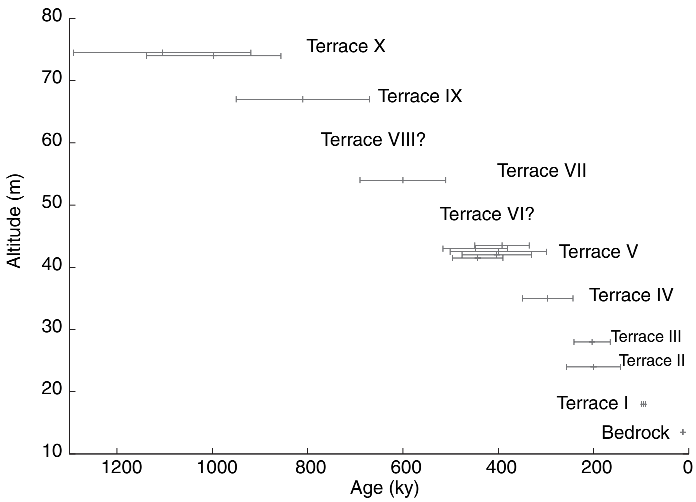

Fig. 4 Age of the fluvial terraces of the Somme (Laurent *et al.*, 1994; Antoine *et al.*, 2000; Van Vliet Lanoe *et al.*, 2000). Elevation of the fluvial records are those observed at Amiens (Antoine *et al.*, 2000).

the year. In conclusion, a 'characteristic' water flux would decrease during warming periods (from glacial to interglacial periods) and increase during cooling periods (from interglacial to glacial periods). This is in agreement with Rotnicki's results (1991), which demonstrate the reduction of the Prosna river water discharge for the last 12 kyr (Poland). The water flux also decreased in NW China over the same time period (Poisson and Avouac, 2004).

If we now check the sediment supply evolution over time, field data suggest that sediment supply increased under colder climatic conditions, due to gelifraction, solifluxion, eolian activity and the diminution of vegetal cover (Antoine, 1994; Antoine *et al.*, 2000). Freeze–thaw cycles weaken the soil and riverbank strength. Under temperate conditions, the vegetation cover is more extended and dense: soils are therefore not easily erodible (Bogaart *et al.*, 2003). Furthermore, we must consider the physical reaction time of vegetation cover to climatic change which is probably different from those of sea-level change, temperature or rain water evolution.

Long-term fluvial erosion model

Various equations have been used for modelling erosion processes: the mass conservation equation (Culling, 1960), the stream power equation (Howard *et al.*, 1994), the universal soil loss equation and the Water Erosion Prediction Project (Nearing and Nicks, 1998).

As the mass conservation equation has been used in numerous studies (Culling, 1960; Begin *et al.*, 1981; Willgoose *et al.*, 1991; Pelletier and Turcotte, 1997; Allen and Densmore, 2000) and is a well-established physical model, we chose to use it as a base for our simulation. There is ample literature on the different applications of this equation and on the analytical and numerical methods to solve it (Carslow and Jaeger, 1959). The theoretical basis of the continuity equation offers the ability to study the influence of the second-order processes such as the lateral mass supply (Begin *et al.*, 1981), the relationship between tectonics and network distribution (Willgoose *et al.*, 1991), the autocyclic dynamics in fluvial sedimentary basin (Pelletier and Turcotte, 1997), the variations of sediment flow in response to the climatic variations (Veldkamp and van Djike, 2000) and the relationship between flexural isostasy and deep incision (Gargani, 2004b).

The mass conservation equation can be written:

$$\frac{\delta z(x,t)}{\delta t} = \frac{\delta q_s(x,t)}{\delta x} + B(x,t) \quad (3)$$

where q_s is the sediment flux per unit width, z the altitude of the river bedrock, t the time, x a longitudinal distance and $B(x,t)$ is the lateral mass supply.

One of the main questions in modelling long-term river evolution is the way to simulate the sediment flux q_s. Various forms of the sediment flux have been tested for the modelling of erosion. Each form seems to be appropriate to the simulation of specific features of rivers and hillslopes.

The simple form of the sediment flux is a consequence of the observation that sediment flux is proportional to the slope S:

$$q_s(x,t) = k_1 S. \tag{4}$$

This hypothesis has often been used not only to model hillslope (Willgoose *et al.*, 1991), but also to model river erosion (Begin *et al.*, 1981).

Nevertheless this method is not appropriate here because it does not take the influence of the water flux on river evolution into account. This approach cannot simulate precisely the processes controlling erosion and cannot describe the river evolution when erosion is limited by the sediment transport capacity or limited by the abrasion capacity of the flow.

Three other hypotheses have been offered for sediment flux modelling. We will now briefly present these hypotheses and their consequences. The first takes into account the impact of the water flux by unit width q on fluvial erosion, in addition to the slope (Willgoose *et al.*, 1991):

$$q_s(x,t) = k_2 q(x,t) S. \tag{5}$$

Under this hypothesis, the mass conservation equation can be written:

$$\frac{\delta z(x,t)}{\delta t} = k_2 \left[q(x,t) \frac{\delta^2 z(x,t)}{\delta x^2} + \frac{\delta q(x,t)}{\delta x} \frac{\delta z(x,t)}{\delta x} \right] + B(x,t). \tag{6}$$

The second considers that the variation over time of the sediment flux is proportional to the difference between the sediment flux q_s and the steady state of the sediment flux q_s^{eq} (Kooi and Beaumont, 1994):

$$\frac{dq_s}{dt} = \frac{1}{t_s}(q_s^{eq} - q_s). \tag{7}$$

This approach allows to obtain an equivalent of the stream power equation (Howard *et al.*, 1994; Montgomery, 1994). Starting with the mass conservation equation, in the case of a detachment limited or undercapacity condition ($q_s << q_s^{eq}$) and considering that the sediment flux at the equilibrium state can be simulated by $q_s^{eq} = k_2 q(x,t) S$, the model can be simply written (Kooi and Beaumont, 1994):

$$\frac{\delta z(x,t)}{\delta t} = k_3 q \frac{\delta z}{\delta x}. \tag{8}$$

This useful approach is unfortunately not appropriate to model large sediment transport, as expected during the glacial times, because sediment flux q_s must be small ($q_s << q_s^{eq}$).

The third assumes that the spatial variation of the sediment flux is proportional to the difference between the steady state of the sediment flux and the sediment flux:

$$\frac{\delta q_s}{\delta x} = k_4 \frac{q_s}{q_s^{eq}} (q_s^{eq} - q_s). \tag{9}$$

This assumption leads to a fluvial erosion model which is able to take into account the consequences of an overcapacity situation of the sediment into the water flux (Sklar and Dietrich, 2001; Van der Beek and Bishop, 2003):

$$\frac{\delta z(x,t)}{\delta t} = k_4 q_s \left(1 - \frac{q_s}{q_s^{eq}}\right). \tag{10}$$

This approach has the advantage of allowing to obtain an equation which is easier to solve, however, under a restrictive and physically unclear condition on the spatial evolution of the sediment flux, without offering an explicit form of the sediment flux.

The erosion model depends on the law adopted for the sediment flux. We will take the model proposed in Eq. (6). This model is able to take into account the slope, the water discharge per unit width, the lithology and the sediment supply variations because of the hypotheses done on the sediment flux. One must also choose how to simulate these parameters. For long-term river erosion, one must take into account the time and spatial variations of these parameters.

Many studies consider only spatial variations without taking into account time variations. They implicitly make the hypothesis that spatial variations are independent from time variations, modelling only the spatial variations for the water flux per unit width for the long term:

$$q(x,t) = q(x)q(t) = \beta q(x) = q(x). \tag{11}$$

We made the assumption that time and spatial variations can be modelled independently for the water flux per unit width and for the sediment supply, these being a characteristic of the fluvial system considered and of the

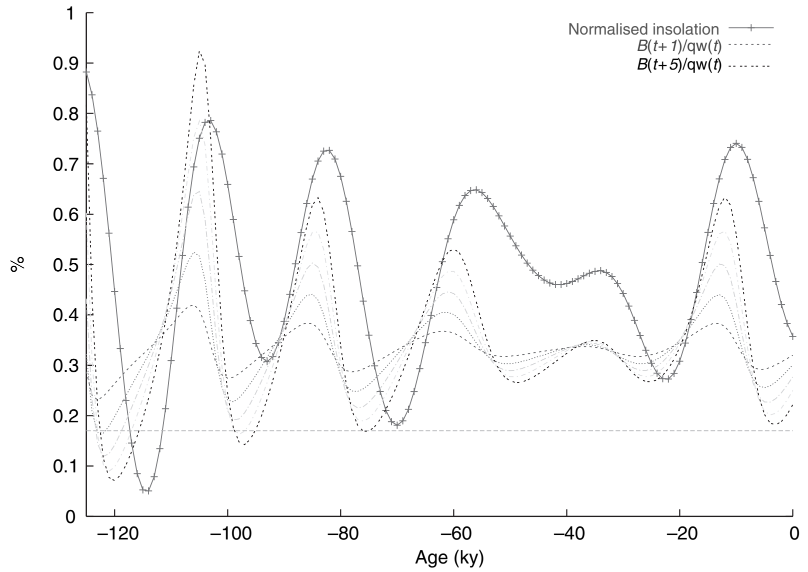

Fig. 5 Influence of the lag of time (1–5 ka) between normalized water discharge per unit width $q(t)$ and lateral sediment flux $B(t)$ on erosion and sedimentation phases. In the example, the spatial threshold, depending on spatial conditions, has been chosen at $-T_s = 0.17$. Erosion will happen only if $B(t)/q(t)$ has a value less important of 0.17. This condition is verified, in this specific example, only for some lag of time between $q(t)$ and $B(t)$ comprised between 3 and 5 ka.

general climatic evolution of NW Europe respectively. We propose to write these fluctuation as the product of two terms, one dependent on space, the other on time:

$$q(x,t) = q(x)q(t) \tag{12}$$

$$B(x,t) = B(x)B(t). \tag{13}$$

$q(t)$ and $B(t)$ can be seen as percentage function oscillating in [0,1]. From Eq. (6) and considering conditions (12) and (13), we obtain:

$$\frac{\delta z(x,t)}{\delta t} = k_2 q(t) \left[q(x)\frac{\delta^2 z(x,t)}{\delta x^2} + \frac{\delta q(x)}{\delta x}\frac{\delta z(x,t)}{\delta x} \right] + B(x)B(t). \tag{14}$$

There is erosion when $\delta z/\delta t < 0$. This is equivalent to:

$$\frac{k_2}{B(x)} \left[q(x)\frac{\delta^2 z(x,t)}{\delta x^2} + \frac{\delta q(x)}{\delta x}\frac{\delta z(x,t)}{\delta x} \right] + \frac{B(t)}{q(t)} < 0 \tag{15}$$

with

$$B(x) > 0$$

$$q(t) > 0.$$

Modelling spatial variations

The term representing the spatial evolution in Eq. (15) is:

$$T_s(x) = \frac{k_2}{B(x)} \left[q(x)\frac{\delta^2 z(x,t)}{\delta x^2} + \frac{\delta q(x)}{\delta x}\frac{\delta z(x,t)}{\delta x} \right]. \tag{16}$$

Erosion depends on the geometrical and hydrological characteristics that are described by this term.

The spatial variations of hydrological parameters can be simulated simply by using standard relations (Montgomery, 1994; Knighton, 1998):

$$Q = a_Q A^{b_Q} \tag{17}$$

$$A = a_A x^{b_A} \tag{18}$$

$$w = a_w Q_b^{b_w} \tag{19}$$

$$Q_b = a_{Qb} Q^{b_{Qb}} \tag{20}$$

$$B = a_B \frac{\delta A}{\delta x} \tag{21}$$

where Q is the water flux, A the drainage area, x the distance to the source, w the river width, Q_b the bankfull discharge, B the sediment supply, and a_Q, b_Q, a_A, b_A, a_B, a_w, b_w, a_{Qb}, b_{Qb} empiricals parameters to be calculated for each specific river.

Using Eqs (17)–(21), we can write the spatial variations of the water flux per unit width $q(x)$:

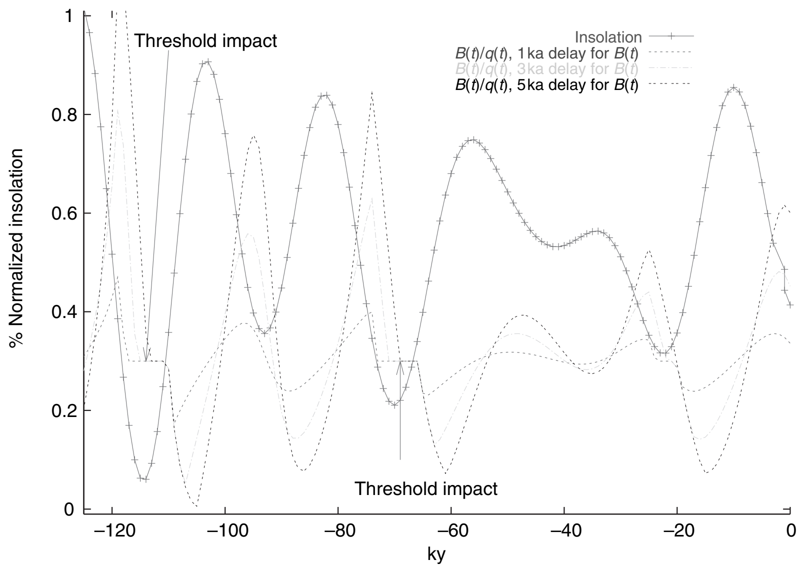

Fig. 6 Influence on time simulation of the thresholds on water discharge per unit width and sediment supply.

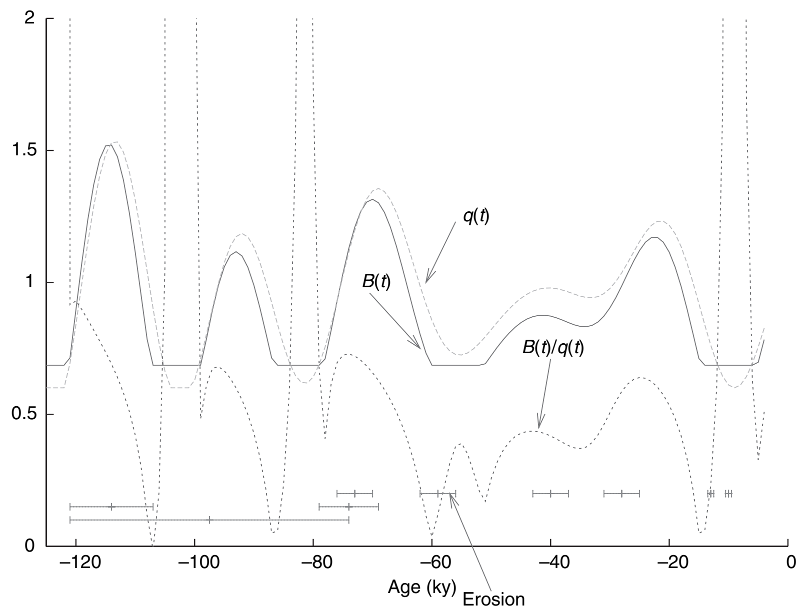

Fig. 7 Influence on time simulation of the thresholds on water discharge per unit width and sediment supply. In the case of large thresholds, time simulation can be modified importantly and produce non-linear effect. The time lag is of 5 kyr. The threshold is 10% of the value of the water discharge.

$$q(x) = \frac{a_Q^{1-b_w b_{Qb}} a_A^{b_Q(1-b_{Qb}b_w)}}{a_w a_{Qb}^{b_w}} \times x^{b_A b_Q(1-b_{Qb}b_w)}$$

and the sediment supply $B(x)$:

$$B(x) = a_A a_B b_A \times x^{b_A - 1}$$

as function of x is necessary to solve Eq. (15).

Modelling climatic variations

We consider that characteristic fluctuations of water flux and sediment supply represent the main features of climatic changes able to modify significantly the river dynamics. We assume that the time lag between water flux and sediment supply is one of the main causes of erosion (Bogaart *et al.*, 2003).

The dependence on climatic changes is described by the term $B(t)/q(t)$ in Eq. (15). This term is always positive. There is erosion only if $B(t)/q(t) < -T_s$ where T_s represents the spatial control defined in Eq. (16) resulting from geometrical and hydrological parameters.

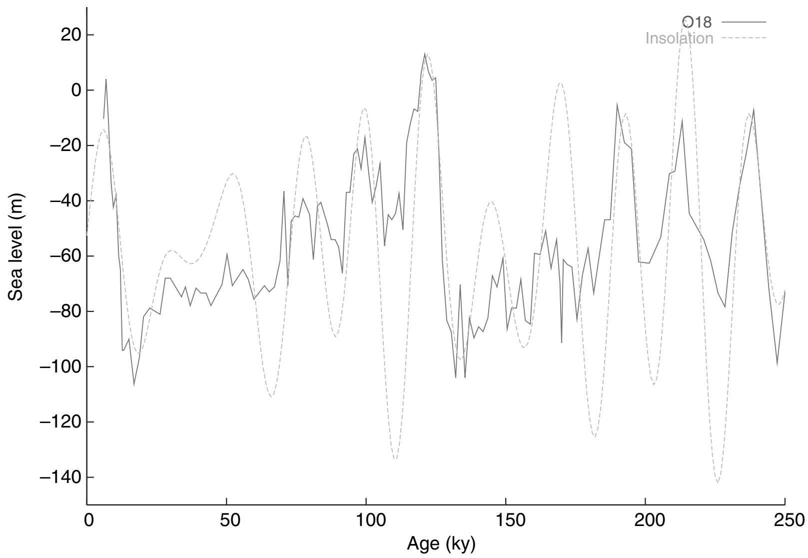

Fig. 8 Simulation of sea-level variation using the insolation curve. Comparison with the ^{18}O proxy. The time lag between the ^{18}O proxy and the insolation curve is of 4 ka.

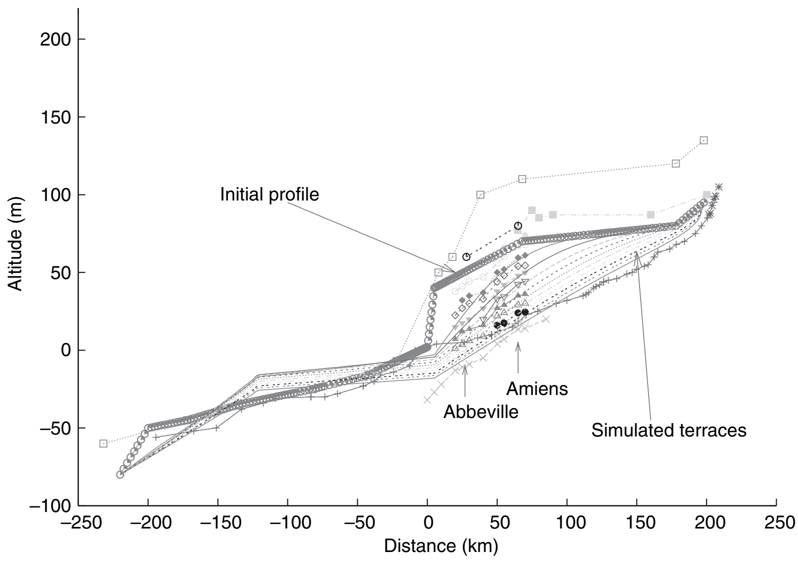

Fig. 9 Hypothesis 1: the initial profile is a plateau and there is no uplift (same as for Fig. 3).

We assume that the characteristic water discharge per unit width and the sediment supply time variations $q(t)$ and $B(t)$ can be simulated using the insolation signal (or alternatively by a δ^{18}O curve). Insolation is the energy delivered from the Sun on Earth due to the variations of the astronomical parameters. It can be considered as one of the main factor influencing climatic change in the long term.

Considering that the characteristic water discharge per unit width and the sediment supply increase during cold periods, and that there are a diminution of these two parameters when climate is warmer, we simulate their evolution using a specific linear transformation for each input parameter of the insolation curve of Berger and Loutre (1991), calculated at 65°N, normalized for the last million years. The variation of the characteristic water discharge per unit width changes in the opposite way to the normalized insolation curve. We calculated the term $B(t)/q(t)$ for different lag of time between characteristic water discharge per unit width and sediment supply (Fig. 5). These numerical experiments show that various periods of time, favourable to erosion or sedimentation happened during the last climatic cycle with a duration and an amplitude depending on the time lag.

The time lag between insolation curve and sea-level change adopted in previous such simulations is 1 kyr (Veldkamp and van Djike, 2000; Veldkamp and Tebbens, 2001). Bogaart and van Balen (2000) considered a time lag of 5 kyr between water discharge and sediment supply. The existence of a time lag is the consequence of the different reaction time for each physical system in response to astronomic change (Shakleton *et al.*, 1990). We assume that changes for the water flux are synchronous with insolation curve and happened before those of sediment supply. Veldkamp and van Djike (2000) also assumed that the insolation may describe the variation of water flux, sediment supply and sea-level change, but assuming another time lag.

It is obvious that the climatic simulation that we propose is a simplification of the many parameters influencing the climatic impact (Fig. 8). Nevertheless, our approach is also able to consider other consequences of the complex impact of climatic change using a threshold in the normalized time variation of $q(t)$ and $B(t)$ (Figs 6 and 7): (1) the fact that under a critical value of the water discharge per unit width no erosion occurs, (2) the fact that a saturation of the protection capacity by the vegetation cover happens for a very high density of vegetation, and (3) the fact that once the lack of vegetation cover occurred under extreme glacial conditions, there was no further increase in hillslope erosion rate (considering the vegetation cover as the main process in hillslope erosion protection).

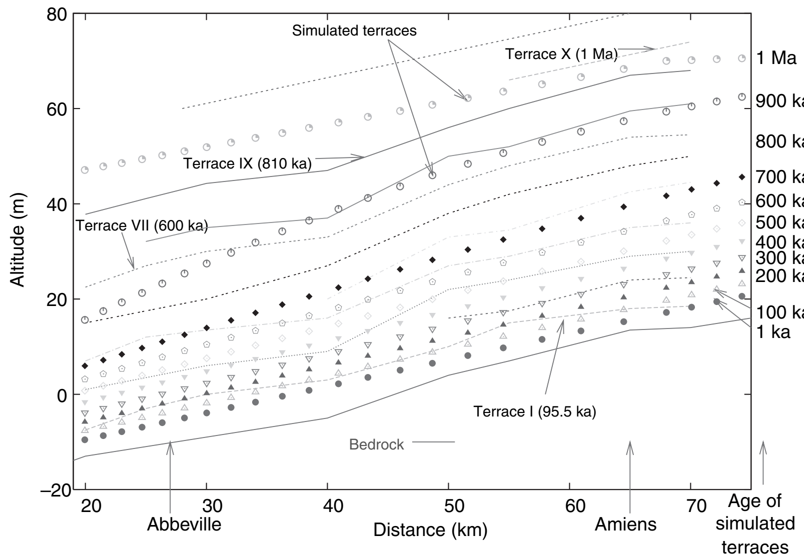

Fig. 10 Hypothesis 1: the initial profile is a plateau and there is no uplift. Lines represent real terraces and modelled terraces are represented by symbols.

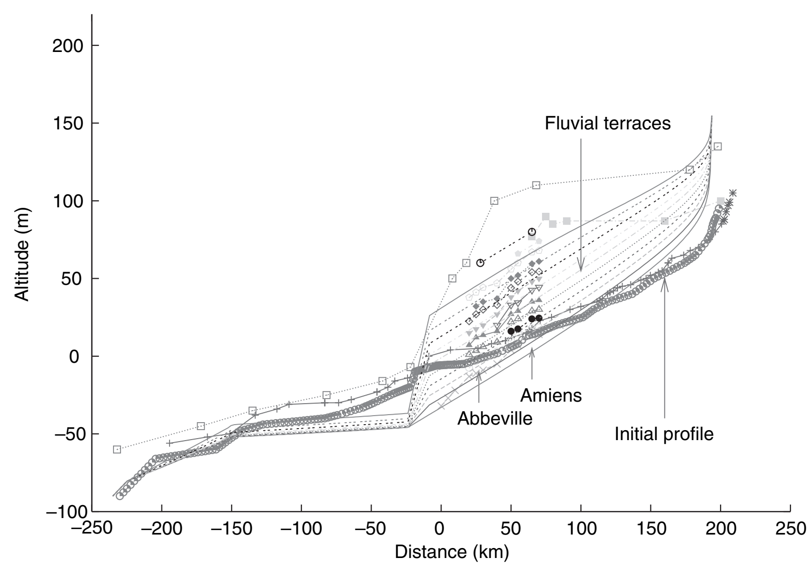

Fig. 11 Hypothesis 2: the initial profile is parallel to the present one and there is 60 m Myr^{-1} of homogeneous uplift in the region upstream to the present coast line (same as for Fig. 3).

Application to the River Somme

The spatial variation of hydrodynamic parameters a_Q, b_Q, a_A, b_A, a_w, b_w, a_{Qb}, b_{Qb} and a_B have been determined specifically for the River Somme in its present state using present hydrological values.

The lithology of the River Somme Valley is homogeneous. We modelled the lithology variations, using the coefficient k_2 of Eq. (6). For Cretaceous chalk and limy facies downstream to the Greenwich deep, we assume that $k_2 = 1000$, whereas for the Tertiary sand and clays we consider that $k_2 = 9000$. The quantification of the coefficient k_2 is arbitrary. Only the relative value of this coefficient between two different lithologies describes the difference of bedrock resistance to fluvial erosion.

Simulation experiments and results

Based on the model presented above, we tested three hypotheses for the initial profile of the River Somme, corresponding to the possible state of the river, 1 million years ago:

1. One million years ago, there was an existing relief (a small plateau). No uplift occurred during the last million years (Fig. 9).
2. One million years ago, the longitudinal profile of the Somme was parallel to the present one. A homogeneous uplift occurred at a rate of 60 m Myr^{-1} for the last million years for the Somme valley, upstream of the present coastline.
3. One million years ago, the longitudinal profile of the Somme was subhorizontal without escarpment. An uplift of 60 m Myr^{-1} took place in the last million years, upstream of the present coastline.

Considering the first hypothesis, the simulated evolution of the longitudinal profile is not compatible with field data because of the high variation of the incision rate during the last million years for the simulated river. Indeed, whereas field terraces indicate a regular incision for the last million years, terraces simulated under this hypothesis predict a rapid incision during the first 300 kyr and were followed by an abrupt decrease of the incision rate of the valley (Figs 9 and 10). Furthermore, the slope of the simulated longitudinal profile is different from the slope of the real fluvial terraces.

Considering the second hypothesis, the evolution in time of the simulated terraces is compatible with field data in the downstream part, however, with a slightly steeper slope than for real terraces (Figs 11 and 12). In the upstream part, there is no published data on fluvial records, so it is impossible to compare our simulation results with field data. Nevertheless, simulated terraces in the upstream part incise in the same proportion to those in the downstream part, whereas the real valley is more incised downstream than upstream.

Considering the third hypothesis, the evolution in time of the simulated longitudinal river profile is also compatible with field data in the downstream part (Figs 13 and 14). We can see that the simulated terraces fit the real terraces. For example, simulated terrace of age *c.* 600 ka feet the real terrace of age 600 ka with an accuracy of ± 10 m. In the upstream part, there is a diminution of the incision of the simulated valley: simulated terraces converge in the upstream part. This simulation satisfies the principal features of the real valley.

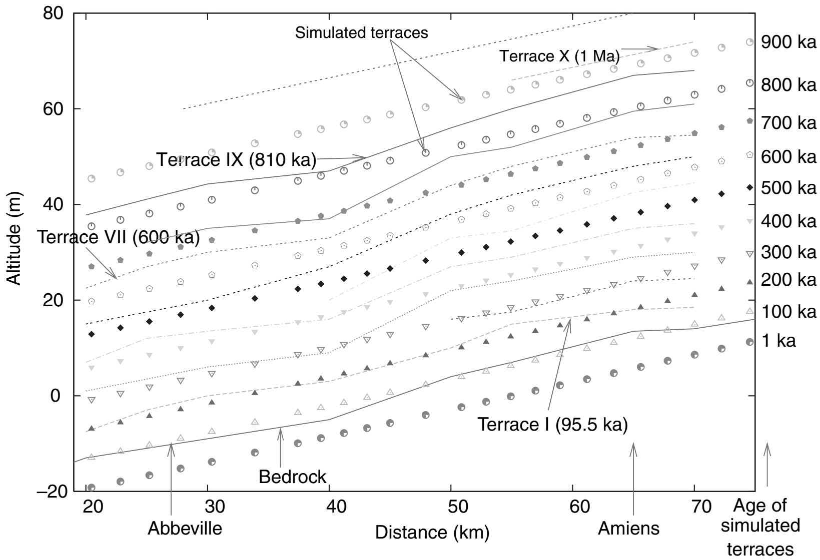

Fig. 12 Hypothesis 2: the initial profile is parallel to the present one and there is 60 m Myr^{-1} of homogeneous uplift in the region upstream to the present coast line. Lines represent real terraces and modelled terraces are represented by symbols.

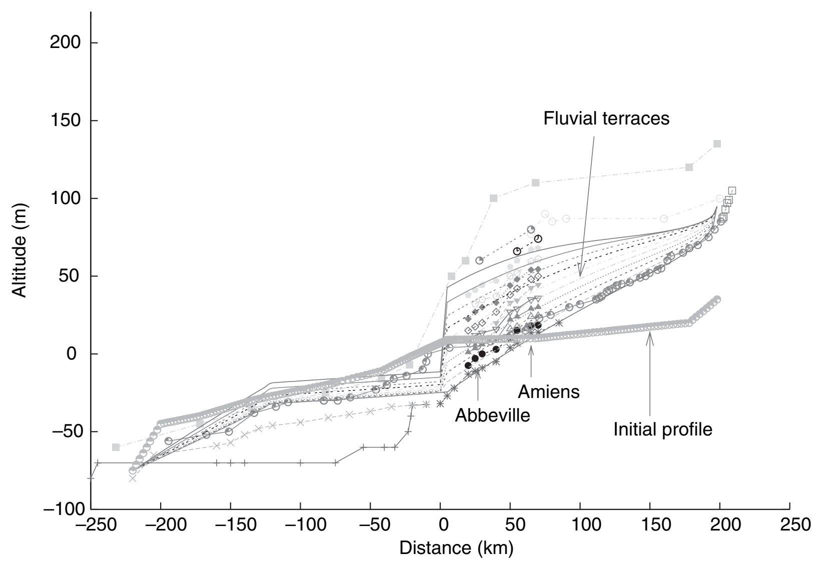

Fig. 13 Hypothesis 3: the initial profile is subhorizontal without initial escarpment and there is 60 m Myr^{-1} of homogeneous uplift in the region upstream to the present coast line (same as for Fig. 3).

Under these initial conditions we studied the simulation in more detail to determine the main periods of bedrock incision. We are also able to offer a definitive age for the maximum incision for the area of Abbeville. For the last climatic cycle, the simulation predicts that fluvial erosion affects principally the bedrock, during the beginning of the climatic cycle in the Abbeville area (Fig. 15). The maximum incision happened between 60 and 40 kyr. After this phase, the conditions for erosion were not satisfied and sedimentation may have affected the river.

Discussion and conclusion

The numerical model presented shows the response of a river profile to climate changes in the case of a moderate uplift for the last climatic cycle. The simulation predicts that erosion occurs, when climatic conditions are able to produce specific hydrological conditions (high water discharge per unit width and limited sediment supply). The time lag between water discharge and sediment supply is a critical mechanism in the process and is certainly one of the major causes of erosion over a timescale of 1–100 kyr for NW Europe fluvial system. The modifications involved by non-linear reaction of river to climate change play an important role, as suggested by the numerical experiments. The simulation predicts that the maximum erosion during the last climatic cycle was between 120 and 60 or 40 kyr due to climate-induced changes and not during the maximum glaciations. This result conforms to the interpretation of field data (Antoine, 1994). Sea-level change has not had a significant influence on the River Somme over these same periods because there is no significant break of slope in the River Somme

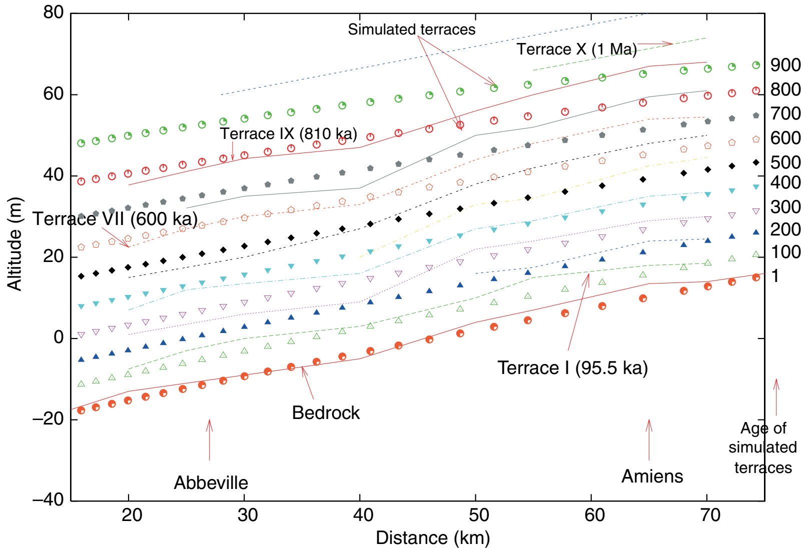

Fig. 14 Hypothesis 3: the initial profile is subhorizontal without initial escarpment and there is 60 m Myr^{-1} of homogeneous uplift in the region upstream to the present coast line. Lines represent real terraces and modelled terraces are represented by symbols.

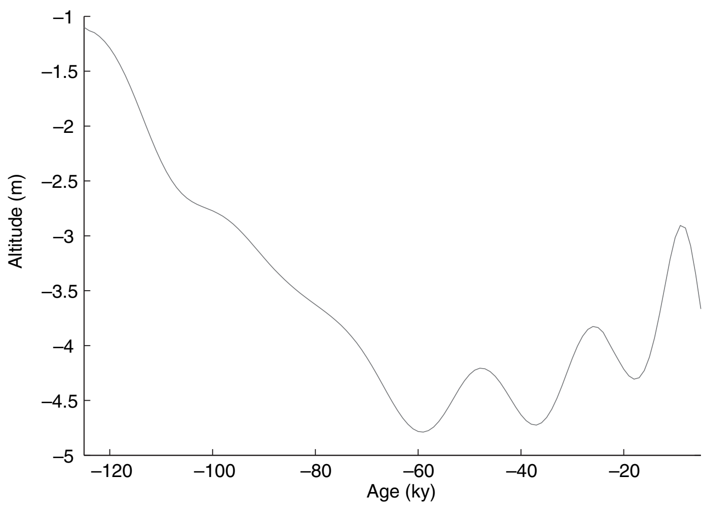

Fig. 15 Simulated incision and aggradation in Abbeville for the last 125 kyr for a tectonic of 60 m MyrEE^{-1}. The simulated maximum of incision during the last climatic cycle happened during the beginning of the climatic cycle, before the glacial phase.

longitudinal profile. Nevertheless, the curve used to model the sea-level change may influence the results in cases where a break of slope exists in the longitudinal river profile.

This simplified model does not take directly into account the possible modifications over time of the drainage area. Nevertheless, in a first approximation, the spatial hydrological features can be considered independent of time because there is no significant modification of the fluvial network such as a capture in the specific case of the River Somme. Improvements to this approach can be made by using more detailed process-based models with more precise and numerous input parameters. The quantification of the time lag has to be studied more precisely. The time lag is certainly not superior to 5 kyr because of the duration of major climatic variations and the possible reaction time for the different physical systems. Nevertheless, there is presently no indication for a precise time lag representative for the last climatic cycle. However, such a detailed modelling approach would require additional data and model applications on a different time-scale, which is beyond the scope of the present paper, but may be the subject of further studies.

On a long time-scale (*c.* 1 Myr), the tectonic uplift is the main control of the incision rate in the Paris Basin. A simulated uplift rate of 60 m Ma^{-1} is compatible with field data for the River Somme. This value is in line with the rates obtained for the Ardennes (Veldkamp and van Djike, 2000) and a little smaller than for the Seine (Gargani, 2004a). Numerical experiments suggest that 1 million years ago, the River Somme certainly had a subhorizontal longitudinal profile, with little relief throughout the Paris Basin.

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