Experimental modelling of tectonics–erosion–sedimentation interactions in compressional, extensional, and strike–slip settings

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ABSTRACT

Tectonically controlled landforms develop morphologic features that provide useful markers to investigate crustal deformation and relief growth dynamics. In this paper, we present results of morphotectonic experiments obtained with an innovative approach combining tectonic and surface processes (erosion, transport, and sedimentation), coupled with accurate model monitoring techniques. This approach allows for a qualitative and quantitative analysis of landscape evolution in response to active deformation in the three end-member geological settings: compression, extension, and strike–slip.

Experimental results outline first that experimental morphologies evolve significantly at a short time scale. Numerous morphologic markers form continuously, but their lifetime is generally short because erosion and sedimentation processes tend to destroy or bury them. For the compressional setting, the formation of terraces above an active thrust appears mainly controlled by narrowing and incision of the main channel through the uplifting hanging-wall and by avulsion of deposits on fan-like bodies. Terrace formation is irregular even under steady tectonic rates and erosional conditions. Terrain deformation analysis allows retrieving the growth history of the structure and the fault slip rate evolution. For the extensional setting, the dynamics of hanging-wall sedimentary filling appears to control the position of the base level, which in turn controls footwall erosion.

Dynamic evolution of topography in tectonically active areas results from complex interactions between deformation and surface processes (erosion, transport, and sedimentation). As a consequence, specific geomorphological, structural, and sedimentary features develop according to the geological context. They are for instance: (i) uplifted or folded terraces, reverse fault or fold scarps, and wind gaps in compressional tectonic settings (e.g., Avouac et al., 1993; Keller et al., 1998; Chen et al., 2007); (ii) triangular facets and wine glass valleys developing along normal faults in extensional settings (Cotton, 1950; Armijo et al., 1986; Gawthorpe and Leeder, 2000); and (iii) offset terraces and channels, beheaded streams, shutter ridges, and sag ponds developing along strike–slip fault zones (Wesson et al., 1975) (Fig. 1). All these depositional or erosional features constitute useful

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geomorphic markers that provide key information to better constrain recent Earth's surface deformation mechanisms and kinematics (Burbank and Anderson, 2001; Keller and Pinter, 2001).

At the seismic time scale, the study of active faults aims at recovering information on past (i.e., last $10^2$–$10^4$ years) large earthquakes to understand the incremental processes accounting for the growth of topography and to improve seismic risk assessment. The classical approach consists in searching deformed or offset markers (e.g., topography, river beds, alluvial fans, drainage networks, terraces), in measuring their shape, and finally in reconstructing their original geometry. Combined with marker dating, it gives quantitative information on active fault parameters, such as earthquake magnitude, time recurrence, clustering, typical failure lengths and amplitude of coseismic surface displacements (e.g., Van Dissen and Berryman, 1996; Van der Woerd et al., 2001; Armijo et al., 2010; Klinger et al., 2011). However, results strongly depend on the recognition and interpretation of chosen deformed markers, whose initial geometry and morphological evolution following their formation and very first fault disruption are often difficult to determine unambiguously.

Tectonic landforms as observed in most mountain ranges generally rise coseismically (e.g., Stein et al., 1988), following a succession of seismic cycles (Reid, 1910). A topographic signal is generated when deformation reaches the surface in the form of a fault or fold scarp. If the seismic cycle duration is shorter than the time required to totally erode the coseismic scarp, the topographic signal accumulates through time and generates a long-term tectonic landform. Linking the short-term component of Earth surface deformation with the long-term cumulative landforms, as recorded in the morphology, is challenging. It requires that landforms contain a detailed and preserved message of their growth history (e.g., Sieh, 1984; Gaudemer et al., 1995; Keller et al., 1998; Manighetti et al., 2001; Pazzaglia and Brandon, 2001; Carretieri et al., 2002; Hubert-Ferrari et al., 2007; Li et al., 2012; Le Bèon et al., 2014; Simoes et al., 2014). In that case, insights can be brought for instance on the steadiness (both in amplitude and direction) of the strain rate or the model of fault break (e.g., characteristic earthquake model). However, this approach faces some difficulties related to the sparse spatial distribution of available data (e.g., from seismics, well logging, geodesy, thermochronology, paleomagnetism, geochemistry) and generally reveals little of past relief evolution.

Questions concerning the formation, evolution, and record of deformation of morphologic markers are still difficult to answer. Similarly, understanding the time scales of landscape responses to tectonic deformation and the spatiotemporal variations of surface fluxes (erosion and sedimentation rates) in relation to tectonic fluxes are worth investigating. To tackle those issues, the development of modelling techniques has contributed to improving our understanding of the feedback mechanisms between deformation and surface processes (e.g., Buiter, 2012; Corti, 2012; Dooley and Schreurs, 2012; Graveleau et al., 2012). Particularly, experimental modelling investigated the response of landforms to changes in internal parameters (i.e., rheology) or external forcing (i.e., tectonics, climate) with various apparatus, length, and time scales: either with the erosion box device (Ouchi, 1985, 2004, 2011; Hasbargen and Paola, 2000; Schumm et al., 2000; Bonnet and Crave, 2003; Lage et al., 2003; Babault et al., 2005, 2007; Turowski et al., 2006; Douglass and Schmeerkecle, 2007; Malverti et al., 2007; Bonnet, 2009; Rohais et al., 2012) or the sand box device (Barrier et al., 2002, 2013; Nalpas et al., 2003; Gestain et al., 2004; Persson et al., 2004; Konstantinovskaia and Malavieille, 2005; Bonnet et al., 2007, 2008; Pichot and Nalpas, 2009; Malavieille, 2010; Malavieille and Konstantinovskaia, 2010; Konstantinovskaia and Malavieille, 2011; Perrin et al., 2013). Within the community of researchers working on relief dynamics, an original approach has been developed in the Experimental Tectonic Laboratory at Geosciences Montpellier, France. Active deformation of experimental materials (including nucleation, reactivation, and propagation of faults) and active morphogenesis (including channel and hillslope processes, transport, and sedimentation) have been closely combined to address the natural mechanisms of topography growth. Experiments either in compression (Graveleau, 2008; Graveleau and Dominguez, 2008), extensional (Strak et al., 2011; Strak, 2012), or strike–slip settings (Chatton et al., 2012) have been performed. This experimental approach enables a survey of the formation and evolution of a fault from its immature stages up to several hundred slip events. The spatial and temporal evolution of fault kinematics and topography is continuously recorded thanks to accurate measurement techniques. Therefore, the impact of erosion and sedimentation on the

Fig. 1. Tectonic geomorphology with relevant morphotectonic markers in (A) thrust fault setting, (B) normal fault setting, and (C) strike–slip fault setting.
fault zone topography can be investigated, and the evolution of associated morphological markers can be described and quantified.

In this paper, we present the morphological evolution of tectonically active experimental landforms covering compressional, extensional, and strike–slip end-member tectonic settings. Our aim is to show through selected striking examples that the experimental approach has a great potential and can be successfully applied to all types of tectonic contexts at a wide range of temporal and spatial scales. We also aim to show that such tools are needed to better determine the limits of quantitative measurements made on actively deforming natural landscapes. In the following, we describe the experimental facilities (devices, experimental material, and scaling) and one typical experiment chosen to illustrate each tectonic setting. Then, a qualitative description of each experiment is used to evaluate the degree of analogy between the morphotectonic evolution of experimental markers and their natural equivalents. Several morphological markers are quantitatively analysed in each setting to illustrate how they form, evolve, and record tectonic deformation. Finally, some challenges and potential improvements are exposed to target future projects.

2. Experimental procedure

2.1. Setup and boundary conditions

The experimental devices developed at Geosciences Montpellier to investigate the morphological evolution of tectonically active landscapes result from laboratory modelling initially focused on convergence zones (Malavieille, 1984; Lallemant et al., 1992, 1994; Calassou et al., 1993; Kukowski et al., 1994, 2002; Chemenda et al., 1995; Gutscher et al., 1996, 1998a, 1998b, 2001; Dominguez et al., 1998b, 2000), but also on wrench settings (Lu and Malavieille, 1994; Dominguez et al., 1998a; Martinez et al., 2002) and extensional contexts (Schlengauff et al., 2008). These sandbox experiments mainly focused on the long-term structural evolution of models without including surface processes (erosion and sedimentation) that were later progressively incorporated to investigate their interactions with tectonics (Malavieille et al., 1993; Larroque et al., 1995; Konstantinovskaia and Malavieille, 2005; Bonnet et al., 2007, 2008; Malavieille and Trullenque, 2009; Malavieille, 2010; Malavieille and Konstantinovskaia, 2010; Konstantinovskaia and Malavieille, 2011; Perrin et al., 2013).

In the last decade, three main types of experimental setups have been designed to investigate the short-term morphological evolution of tectonic landforms taking into account more realistically erosion, transport and sedimentation processes (Fig. 2). These setups share common features like (i) a computerized deformation mechanical apparatus, (ii) a programmable rainfall system composed of a sprinkler network, and (iii) an optical measurement device. Computerized deformation devices are slightly different within the three setups to face the required boundary conditions associated to their tectonic contexts (thrusting, normal faulting, strike–slip), whereas they share the same rainfall and optical measurement systems. For the convergent setting, the deformation device is constituted by a large table (100 cm wide, 2000 cm long) at the base of which a mobile film is overlaid by the experimental material. As the film is pulled beneath a rigid backstop, the analogue material is deformed along the backstop, accretes, and thickens. Relief created during the experiment is simultaneously eroded using the rainfall system that generates water surface runoff (Graveleau and Dominguez, 2008). For the extensional (Strak et al., 2011) and strike–slip boundary conditions (Chatton et al., 2012), a setup derived from a prototype work (Bonnet, 2002; Graveleau, 2004) has been adapted to modulate fault strike and dip. The deformation device is composed of two plates (total size 130 × 100 cm) representing the footwall and the hanging-wall of a normal or reverse fault or the two compartments of a strike–slip fault. In the extension configuration, the footwall is maintained as fixed, while the hanging-wall can move along a predetermined normal fault plane separating the two plates. A rigid or a flexural tray can be used to simulate a uniform or a flexural subsidence of the hanging-wall, respectively. In the strike–slip configuration, two crossed motor-driven linear advances allow us to adjust the slip direction from purely strike–slip to normal or reverse strike–slip. The two rigid plates slide along a vertical fault plane following the imposed slip, allowing the analogue material placed on top of it to be deformed.

Fig. 2. Experimental setups for (A) thrust fault setting, (B) normal fault setting, and (C) strike–slip fault setting. Photography of the device (upper panels), and two-dimensional cross-sectional sketch (lower panels). All apparatus are composed of three elements: (i) a computerized deformation device, (ii) a rainfall system (sprinkler device), and (iii) an optical measurement system (CCD camera and laser interferometer).
The rainfall system is similar from one configuration to the other, although small changes in precipitation rate and distribution are used. The rainfall does not intend to simulate real water droplets but should be considered as a mean to trigger through water runoff at the model surface erosional processes on hillslope and channelized processes in topographic lows. Sprinklers are arranged about 1.5 m above the experiment surface to deliver a homogeneous amount of water (in the range of 25 ± 5 mm/h), continuously or during chosen rainfall cycles.

Water droplets are small enough to prevent significant splash erosion but large enough to contribute to small-scale hillslope processes (Graveleau et al., 2011).

Continuous monitoring of the model is performed with the combined use of a laser interferometer and several CCD cameras. Digital pictures shot regularly during the experiment allow for a high-resolution record of the morphological evolution of the model surface. The use of accurate technique of subpixel spectral image correlation derived from satellite imagery processing (Van Puymbroeck et al., 2000; Michel and Avouac, 2002; Dominguez et al., 2003; Avouac et al., 2006; Graveleau et al., 2014) allows us to measure the horizontal displacements over the whole model surface, mapping in detail fault geometry and quantifying surface deformation, particularly fault kinematics. In addition, regular stops of the rainfall device allow us to monitor topography through the generation of digital elevation models (DEMs) with an average spatial resolution of 0.5 mm and a submillimetric vertical accuracy. Finally, models are cut in serial cross sections at the end of the experiment to study the three-dimensional geometry of faults and syntectonic deposits.

2.2. Material

One of the central points of this experimental approach is the development of a specific granular material (MatIV) to model simultaneously deforming processes (essentially faulting) and surface processes (erosion, transport, and sedimentation) as observed in natural landscapes (Graveleau et al., 2011). Its composition has been determined using empirical and physical criteria derived from the comparison of deformation, erosion—transport, and sedimentation processes between nature and models (Graveleau, 2003, 2004, 2008). First, the material satisfies Mohr-Coulomb failure criterion and generates brittle deformation (faults and shears bands) in response to tectonic stresses. Second, it erodes by channel incision and hillslope mass-wasting processes when submitted to water runoff. These processes trigger the development of individualized drainage basins limited by crest lines. Finally, material grain size and shape variability promote particle sorting during transport and trigger stratified sedimentation.

The composition of the experimental material is made up with four granular components (silica powder, glass microbeads, plastic powder, and graphite; Table 1) mixed with water. It has slightly evolved along experimental works to improve the similarity of morphotectonic and sedimentary markers with nature. Notably, slight changes regarding the proportion of silica powder versus plastic powder have been made to enhance hillslope processes versus river incision (Strak et al., 2011). Nevertheless, the physical properties of the wet granular mixture are very similar for all experiments, with an average water saturation rate of 23 ± 3%, a median grain size of 90 ± 10 μm, a porosity of 34 ± 1%, a permeability of 4.9 × 10⁻¹³ m² (0.5 Darcy), a bulk density of 1.61 ± 0.1 g/cm³, a dynamic-stable angle of internal friction of 37 ± 3° and a dynamic-stable cohesion of 675 ± 75 Pa. These physical properties allow for a realistic modelling of morphotectonic and stratigraphic features, including channel network with drainage basins, hillslopes and channels, fluvial terraces (strath or alluvial), knickpoints, alluvial fans with stratigraphic markers, faceted spurs, fault scarps, and pressure ridges (Fig. 3). All these markers display striking morphological similarities with their natural counterpart that provide a validation of the approach.

2.2.1. Scaling

Previous experimental works demonstrated that although scaling works fairly well in tectonic sandbox modelling (Hubbert, 1937;
Horsfield, 1977; Davy and Cobbold, 1991), it is a more difficult goal to achieve in morphodynamic models (Peakall et al., 1996; Paola et al., 2009). The main reasons are notably linked to the inappropriate use of water as erosion–transport agent and to the coexistence of several orders of magnitude for time-dependant processes in nature, which makes direct downscaling in a model challenging. However, scaling in

**Thrusted fault setting**

- Erosion surface
- Channel
- Cumulative scarp
- Incremental scarp

**Normal fault setting**

- Crest line
- Fault trace
- Faceted spur

**Strike-slip fault setting**

- Pressure ridge

**Alluvial fan**

- Proximal facies
- Sorting
- Distal facies

**Fig. 3.** Examples of morphologic features obtained in the three experimental boundary conditions illustrating the experimental material ability to model some natural geomorphic features of tectonically active landscapes.
geomorphic experiments respects the most important force balances for hydraulics (Niemann and Hasbargen, 2005; Malverti et al., 2008), which makes them of value in comparing experiments to long-term dynamics and behaviour of natural systems. As a consequence, to extrapolate quantitative results in the experiments toward nature, we generally consider an average length ratio \( L^* \) (length in model divided by length in nature) ranging from \( 2.1 \times 10^{-4} \) to \( 2.1 \times 10^{-5} \) (1 cm in the model equals 50 to 500 m in nature), depending on the spatial scale of the considered experimental model (from a single fault to a mountain front). The scaling is achieved by comparing qualitatively and quantitatively experimental landforms to their natural counterparts through morphometric parameters (Graveleau et al., 2011; Strak et al., 2011). For the same reasons, the time ratio \( T^* \) (time in model divided by time in nature) ranges from \( 10^{-10} \) to \( 10^{-10} \) (1 s in the model equals about 30 to 300 years in nature), depending on the temporal scale of the investigated issue (from a 100-year seismic time scale to a 100-ky mountain range front evolution). These values are estimated by comparing average erosion rates in models and nature (Graveleau et al., 2011; Strak et al., 2011). For both ratios, we insist on the fact that they give a rough geometric and temporal scaling relationship between the models and their natural counterparts. Therefore, the times, distances, and rates in the experiment converted to times, distances and rates in a natural setting in the following sections (text and figures) should be taken carefully. For instance, calculation of scaled time and distance in the extensional and compressional sections are made with \( L^* = 2.1 \times 10^{-5} \) and \( T^* = 1.5 \times 10^{-10} \) because models are set to simulate topography evolution at the scale of a mountain range. The scaling ratios allow us to give rough estimates that we consider useful to provide to geologists and should not be taken as accurate calculations.

2.3. Reproducibility

Reproducibility of landscape formation and evolution has been tested with multiple runs of a given experiment in the extensional and compressional settings. For instance, with identical rainfall rate and fault slip rate, the results indicate a good reproducibility, with production of the same number of triangular facets and watersheds, the same relief amplitude, and the same values of river and hillslope gradients. Reproducibility has not been intensively tested for the strike–slip setting but similar results can be reasonably expected.

3. Results

3.1. Compressional tectonic context

Morphotectonic experiments coupling tectonic, erosion, and sedimentation processes in compressional tectonic contexts were first presented in a paper focused on the evolution of a fold-and-thrust belt front (Graveleau and Dominguez, 2008). They showed how the propagation of deformation, and notably the nucleation of a frontal thrust, influenced the dynamics of the channel network and associated alluvial deposits. We present, hereafter, an example extracted from the experiment discussed in Graveleau et al. (2011) (Fig. 4), illustrating how channels and sediment deposits evolve in response to fault nucleation and constant hanging-wall uplift. The frontal thrust (thrust 2) is located about 60 cm (scaling to ~30 km in nature) in front of a major hinterland structure (thrust 1; Figs. 4 and 5). The kinematic analysis of images throughout the investigated period of time indicates that convergence is accommodated on thrusts 1 and 2 at constant but different rates (Graveleau, 2008). Shortening rate accommodated on thrust 1 is \( V_{\text{thrust 1}} = 9 \text{ mm/h} \) (scaling to ~0.6 mm/y in nature) and \( V_{\text{thrust 2}} = 26 \text{ mm/h} \) (scaling to ~1.7 mm/y in nature) on thrust 2. Therefore, the frontal thrust accommodates 75% of the convergence. The subsurface geometry revealed on the final cross section indicates that the frontal thrust can be described through a simple fault-bend fold model (Suppe, 1983) with an average ramp dip of \( \alpha = 35 \pm 2^\circ \). We can therefore consider that hanging wall topography has grown at a constant rate of \( U = V \tan \alpha = 18.2 \pm 1.4 \text{ mm/h} \) (scaling to ~1.2 mm/y in nature).

3.1.1. Morphotectonic evolution along a thrust fault

A first movie (online material A.1) shows the morphotectonic evolution of the model and particularly the evolution of the cumulative topography along thrust 1, which cumulates about 12 cm of slip (scaling to ~6 km). As described in Graveleau and Dominguez (2008), such a movie enables us to understand how the nucleation of thrust 1 rapidly triggers uplift of the hanging wall, then its progressive dissection by the development of parallel watersheds, and the deposit of large lobed alluvial bodies in the footwall of the thrust. During this sequence of topographic evolution, erosion and transport processes are responsible for the incision and widening of the channels and the headward propagation of catchments. Deposition of eroded particles occurs temporarily within the channel network and permanently in the foreland. Finally,
large and fast landsliding structures develop along hillslopes of the active topographic front, a feature observed in nature (e.g., Pinto et al., 2008). The evolution of drainage network through time indicates a reduction of catchment number and an increase in drainage basin area (Graveleau, 2008), a feature to be related with the regular spacing ratio of transverse drainages in active topography (e.g., Hovius, 1996; Talling et al., 1997; Castelltort and Simpson, 2006).

In the following, we detail the sequence of morphotectonic evolution of thrust 2 fault scarp (Fig. 5), which has undergone much less cumulative slip (about 4 cm, scaling to ~2 km in nature) than thrust 1. The evolution of thrust 2 being partially framed in movie A.1, its morphotectonic evolution is presented as a separate movie shot in two-dimensional map view (supporting online material A2). The analysis starts after the nucleation of thrust 2 (Fig. 5B), at about 60 min of experimental time (scaling to ~720 ky in nature). This step represents 45% of experimental model bulk shortening. A slight component of folding prior to faulting occurs beforehand, but it is minor and not represented here (see Bernard et al., 2007, for a description of this folding stage). As deformation remains localised on this frontal thrust, the accommodation of shortening induces the uplift of the hanging wall, which is underlined by the creation of a thrust fault scarp. Thrust fault trace is slightly curved and displays several inflexions, which could likely follow local heterogeneities in the material strength. The topographic scarp generates a local increase in surface slope, which, once greater than the erosion threshold of the experimental material, is responsible for the initiation of incision and the development of a channel network. The geometry of the channels starts first as single-threads but it rapidly splits upstream into a ramified pattern. The upstream growth of catchments occurs through headward erosion. Material particles coming from the erosion of the topographic scarp are transferred downstream through the drainage system and accumulate on fan-shaped deposits (S1) similar to natural alluvial surfaces. At the next step (Fig. 5C; 80 min scaling to ~960 ky), a new channel and associated alluvial fan develops in the centre of the studied frame. This channel partly captures the water flux from the rightward watershed, which starts to starve its alluvial body (S2). From this step, a major alluvial system develops in the centre of the frame, and the drainage system to the right progressively dies out (Fig. 5D – F). Meanwhile, channels start to narrow their active bed, incise their substratum and abandon flights of terraces. A series of erosion terraces is identified (T1, in blue; T2, in purple; T3 in red and T4, in orange) on the central and right drainage basins. The central channel being very active, it illustrates an interesting history of channel incision, terrace abandonment and alluvial fan deposition. Indeed, deposition on the alluvial fan occurs on the right side (surfaces S3 and S4), and the channel abandons a first level of terrace on its left bank (T1). Then the channel avulses leftward and deposits transported particles partly over the previous ones (surface S5) and partly in the left lowlands. At that time, channel flow is divided into two flumes, which create an island-shape terrace surface (T2). Avulsion continues and finally stops when the channel blocks leftward at an inflexion of the surface rupture. In the meantime, two additional levels of terraces are abandoned (T3 and T4) on the left bank, creating a step-like pattern of incision terraces. It is worth noting here that terraces have formed irregularly through time, although tectonic fluxes and precipitation condition were held constant.

3.1.2. Quantification of terrace deformation

During the experiment, four digital elevation models (DEM 1, DEM 2, DEM 3, and DEM 4) have been created to quantify the surface evolution of the model (Fig. 6). They correspond to a cumulated shortening of 7, 23, 51, and 58 mm relative to the initiation of the frontal structure, respectively. The morphology corresponding to DEMs 2, 3, and 4 are shown in Fig. 5B, D and F, respectively. This series of four DEMs allows us to compare (i) the continuous growth of topography created above the active frontal thrust fault, and (ii) the deformation recorded by abandoned and deformed terraces. With this aim, we extracted (i) for each DEM, a topographic profile at a fixed place to the left of the alluvial fan area (thick black profile on Fig. 6A) and compared its deformation pattern with (ii) the total finite deformation recorded by the flights of deformed terraces T1, T3 and T4 at the final stage of the experiment (Fig. 6A, C, D). Successive topographic profiles represent the deformation of a pristine surface that was not affected by erosion. The T2 island-shaped terrace is not investigated because its surface is too small to extract reliable topographic measurements.

Results reveal first that the growth of topography in the non eroded area generates a folded surface (Fig. 6B). Maximum elevation culminates above the frontal lowlands at ~5 ± 1 mm for DEM 1, ~20 ± 1 mm for DEM 2, ~34 ± 1 mm for DEM 3, and ~40 ± 1 mm for the final DEM 4. This is in good agreement with calculated total uplift from cumulated shortening (5, 16, 36, and 40 ± 1 mm, respectively), if we assume that the thrust dip (~35 ± 2°) was constant throughout the experiment (Table 2). As a consequence, considering the age of each surface, which is 15, 60, 100, and 130 ± 1 min, respectively (Table 2), an average uplift rate of 18.7 ± 1.0 mm/h can be calculated (scaling to ~1.2 mm/y in nature). This is in agreement with the value calculated from average thrust dip and shortening data (i.e., 18.2 ± 1.4 mm/h).

In addition, we have extracted the topographic profile of each terrace (in orange, red, and blue for T1, T3, and T4, respectively) and have compared their geometry relative to the channel bed (RB) (Fig. 6A, C, D). Raw topographic data are projected along a N25° trend, which is perpendicular to the average strike of the thrust fault (~N115°) at the channel position. A virtual “north” is considered here upward. The active channel bed profile displays a sigmoid shape with a concave upward domain downstream and a convex upward domain upstream (Fig. 6D). The increase in local slope of the channel bed upstream of the fault is directly a consequence of active uplift, as observed in several natural orogenic rivers (e.g., Seebir and Gornitz, 1983; Ouchi, 1985; Schumm et al., 1987). With larger slope, the stream power of the channel is increased, which allows incising through the active structure. Regarding terrace morphology, old terraces are topographically higher than younger ones because they record a longer uplift period. Terrace gradient, notably for T1, is parallel to the present active channel bed, which means that the paleochannel bed has been uplifted without tilting. This is in agreement with the fault-bend fold style of the structure and the expected uplift pattern (Hubert-Ferrari et al., 2007; Simese et al., 2007). Terrace deformation in the field is generally calculated relative to a hypothetical steady riverbed by subtracting the riverbed-long profile from the terrace profiles (e.g., Molnar et al., 1994; Poisson and Avouac, 2004). In our model, we have measured null relief (~0.5 ± 1.0 mm) for T4, ~1 ± 1 mm for T3, and ~5 ± 2 mm for T1, relative to the channel bed (Fig. 6D). Although these measurements are very small and contain relatively large error bars, it is possible to calculate a mean uplift rate by considering the moment when the active channel abandoned them (0 min for T4, 5 min for T3, 11 min for T2, and 26 min for T1) and the stable vertical position of the channel during the experiment. Calculation yields an average uplift rate of 1.15 ± 0.6 mm/h, which is acceptable with the imposed value considering the error bars. Note that another investigation of terrace deformation was studied in Graveleau (2008) on a much bigger thrust system uplifting a set of eight terraces (see Fig. V.42 and 43, p. 422 in Graveleau, 2008) and yielded calculated uplift rates that are consistent with imposed values.

3.1.3. Quantification of erosion and sedimentation amount

At the end of the experiment described in Figs. 4 to 6, the tectonic deformation is stopped during 25 min (scaling to ~300 ky in nature).
but surface processes are maintained active. Therefore, rainfall precipitation continues to trigger channel incision and sediment transport during that period, and sediment deposit keeps going on alluvial surfaces. Consequently, this experimental phase simulates the relaxation of a tectonically driven topography during a phase of deformation quiescence. Such a phase of tectonic quiescence is generally called topographic relaxation (e.g., Lague et al., 2003). To quantify the evolution of surface flux during this period, we have computed a map of erosion and sedimentation amounts by subtracting the two DEMs made at the end of the tectonic and quiescence phase (Fig. 7). Results indicate heterogeneous topographic changes across the model surface, ranging from $-2\text{ mm}$ (erosion) to $+2\text{ mm}$ (aggradation). Maximum erosion and sedimentation rates are therefore about $4\text{ mm/h}$ (scaling to $0.3\text{ mm/yr}$ in nature). Higher amounts of erosion (in blue) are observed within drainage basins, whereas flat surfaces on top of the structure are preserved (in white). Channel banks are particularly eroded, meaning that channels widen. Note that former investigated terraces T3 and T4 have been eroded while the downstream part of T1 still persists. In the main channel, incision is enhanced upstream, meaning that the bulk gradient of the channel-long profile is reduced. Sedimentation (in red) occurs in the central area of the fan meaning that it is backfilling rather than prograding (Schumm et al., 1987; DeCelles et al., 1991). This contributes also to the bulk channel gradient decrease observed upstream.

Table 2
Characteristics of the topographic envelope surface of the frontal structures; calculated uplift are obtained considering an average thrust dip of $35\pm2^\circ$; all cumulated shortening, calculated uplift, measured uplift, and age are relative to the nucleation of the frontal structure.

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Fig. 6. Terrace deformation above an active thrust. (A) Final stage digital elevation model (DEM) showing the location of investigated topographic profiles. ‘Env.’ is for envelope profile reported in (B); RB is for channel bed; T1, T2, and T3 in blue, red, and orange, respectively, are terraces reported in (C) and (D). Topographic profiles in (D) are elevation data projected along a trend ($\pm25^\circ$) perpendicular to the average strike of the thrust fault (about N115$^\circ$) at the channel position; considering that a virtual ‘North’ is upward. Vertical exaggeration is $>2.7$ for (B) and $>3.4$ for (D).
3.1.4. Subsurface geometry of thrust fault and syntectonic sediments

The three-dimensional subsurface geometry of the thrust fault and syntectonic deposits in the footwall basin has been imaged at the final stage of the experiment by sectioning the model. We do not present cross sections through the frontal thrust (thrust 2) because available pictures were not as illustrative as those for the rear thrust (thrust 1) (Fig. 8). The cross sections allow us to follow the geometry of the thrust plane from the deep décollement level up to the surface (Fig. 8A). The transversal fault plane geometry presents three changes in dip from the deepest levels (~45° then 30°) to the shallowest levels (~45° then 25–30°). These changes in dip are most probably linked to the three changes in material rheology from bottom to top, which are the MatII/MatIV, MatIV/stratified sequence, and stratified sequence/syntectonic deposits transitions. The longitudinal geometry of the fault plane is displayed in its shallowest portion (Fig. 8B), where topography, thrust plane and alluvial deposits interact. Because of the gentle dip of the thrust near the surface, the classic rule of V’s is respected as the surface trace of the fault makes a V pointing upstream (in red dashed line) in the direction of the thrust dip. Interestingly, the subsurface geometry of the fault plane appears slightly wavy, as the fault is deeper underneath the drainage basin interfluvies than underneath the channels. Undulation of the fault plane is highlighted by corrugation of the graphite reference level and microbead layers in the stratified sequence that were originally subhorizontal. Here, we do not describe this undulation in terms of folding because it does not result from along-strike horizontal stresses. Instead, we consider it as resulting from the overthrusting of a plastic thrust sheet over a nonplanar sedimentary surface. This sedimentary surface presents an undulated topography owing to the lobed geometry of the syntectonic alluvial bodies. More work on this topic would be needed to investigate if this influence of foreland
sedimentation on thrust fault geometry is relevant or not in nature. Finally, several tectonic slices are observed on the right portion of the longitudinal cross section. Most of them appearing in the strike of major channels and alluvial depocenters, we consider as illustrating a typical break-back sequence of thrusting caused by locally large sedimentation rates (Barrier et al., 2002, 2013).

In addition, syntectonic deposits can be investigated on the final cross sections to correlate the geometry of bedding with the known topographic, tectonic and erosion/sedimentation histories (Figs. 3B and 8B). In our example, an upper layer of unconformable sediments seals the thrust along the longitudinal cross section and masks its surface trace. The surface lying at the base of the unconformable sediments marks the stop in tectonic deformation during the phase of topographic relaxation while erosion–sedimentation was still ongoing (see Section 3.1.3).

3.2. Extensional tectonic context

Laboratory morphotectonic experiments also have been applied to the study of normal fault scarp evolution to investigate the role of fault slip rate on morphology dynamics (Strak et al., 2011). To illustrate the main contributions of this approach, we report on one experiment characterized by a constant fault slip rate of ~2.3 cm/h (scaling to ~1.5 mm/y in nature) and a 26 ± 4 mm/h rainfall rate. The cumulated fault slip was homogeneous along the normal fault and was simulated by mechanically imposing a flexural subsidence of the hanging-wall. The fault dip angle was 60°, allowing us therefore to create a tectonic topography at a rate of ~2 cm/h (scaling to ~1.3 mm/y in nature). These parameters insured a good similarity with the morphotectonic markers observed along normal faults (i.e., best preserved triangular facets; Fig. 9). The morphotectonic evolution of the model surface is featured in supporting online material A3. The precise sequence of morphotectonic evolution is described in Fig. 10 and presented as an additional movie in online material A4.

3.2.1. Morphotectonic evolution along a normal fault

Soon after the beginning of the experiment, a dense hydrographic network with short fault-perpendicular gullies developed on the footwall side of the fault (movies A3 and A4). These incipient gullies rapidly coalesce to form major fault-perpendicular incising channels and associated watersheds (Fig. 10A and B). Hillslope processes dominate between the major channels allowing for the formation of trapezoidal facets, which are delimited by two oblique crest lines trending a few tens of degrees (20° on average) apart from the main crest line direction. On the hanging-wall, sorted particles deposit in alluvial fan-like bodies downstream in the major channels and in the form of interbedded, colluvial and alluvial deposits down the trapezoidal facets. After an experimental time of ~45 min (scaling to ~0.6 My in nature), a stable number of trapezoidal/triangular facets and watersheds is reached (Fig. 10B). The major incising channels continue elongating upstream by headward erosion, which in turn results in erosion of the watershed slopes by hillslope processes and incising gullies, and leads to the progressive dissection of the initially flat topography (Fig. 10B-F). In response to continuous eroded material feeding, alluvial fans prograde in the subsiding hanging-wall (Fig. 10B-F) and elongate downstream inducing the migration of the fault-parallel river that defines the base level. Geometry of the trapezoidal facets also evolves progressively, becoming triangular when their two lateral crest lines merge upfault (Fig. 10C and D). Facet surface is eroded mainly by mass wasting processes, which induce a rapid erosion of the exhumed fault plane. As a consequence, their shape in cross section appears concave upward. Lateral facet edges are best preserved allowing us to document their geometric evolution.

3.2.2. Geometrical evolution of triangular facets and syntectonic sedimentation

The height of the triangular facets increases continuously during the experiment and reaches an almost constant value of ~2.0–3.5 cm (scaling to ~1000–1750 m) at the final stage (Figs. 3C and 10F). The width of the triangular facets is ~2.5–8.0 cm (scaling to ~1250–4000 m) and is stable from the experimental time of ~60 min (scaling to ~0.8 My). These results suggest that a topographic equilibrium is reached close to the fault only at the end of the experiment, despite the apparent steady along-fault morphology (e.g., constant watershed spacing and width of triangular facets) reached earlier.

The geometry of syntectonic deposits in the hanging-wall basin shows a downstream progradation (Figs. 3D and 11A). A fault-perpendicular sorting of deposits is observed, with the relatively denser particles (glass microbeads and silica powder) being deposited closer to the fault scarp than the lighter particles (plastic and graphite powders;
Figs. 3D and 11A). There is also an along-fault sorting (lateral facies transition) of particles with the relatively lighter grains preferentially deposited at the base of the trapezoidal/triangular facets and the denser ones at the outlet of the major channels (Fig. 11B). This indicates that erosion–transport processes are less energetic on and at the base of the trapezoidal/triangular facets than those occurring in the major incising channels and on the alluvial fans. The along-fault geometry of the deposits in the hanging-wall basin indicates moreover that avulsion is a dominant process in and down the major incising channels (see the movie in supporting online material A.4).

3.2.3. Evolution of channel-long and crest-long profiles

During the experiment, eight DEMs were generated every 30 min (scaling to ~0.4 My) (Table 3) allowing us to produce topographic profiles. The evolution of one channel-long profile (blue line in Fig. 9) and one of its adjacent crest-long profile (green line in Fig. 9) is studied considering either the pristine surface of the footwall (Fig. 12A and C) or the base level (Fig. 12B and D) as a reference for elevation. In the first case, the base level falls and migrates continuously away from the fault, controlling the slope of the hanging-wall deposits and the slope of the channel-long (Fig. 12A) and crest-long profiles (Fig. 12C). During the first half-duration of the experiment (i.e., until DEM#5, in green, at ~150 min scaling to ~2 My), these slopes increase with progressive progradation of sediments filling the hanging-wall basin. The slope and the height of the triangular facets also progressively increase (Fig. 12C). In the case where the base level is set as a reference for elevation, a topographic equilibrium is reached close to the fault in the hanging-wall and in the footwall from the experimental time of ~150 min (DEM #5, in green). It is characterised by a topographic slope that remains constant through time (Fig. 12B and D). At the surface of the alluvial deposits, the equilibrium profile elongates as the base level migrates with the progradation of sediments. The equilibrium profile develops up to ~10–15 cm upfault for the major incising channel and up to ~7 cm upfault for the crest line, showing that the topographic equilibrium is reached first in the channels and propagates faster upstream in the channels than on the hillslopes. This observation indicates that the topographic equilibrium is not reached over the entire model surface. Therefore, our experiment illustrates a case of growth stage of the relief along a normal fault. During the phase of upstream propagation of the topographic equilibrium, the slope of the triangular facets reaches its threshold value while their height progressively increases until reaching a constant value at the end of the experiment (Fig. 12D).

3.2.4. Quantification of erosion

Erosion rates were measured between two successive DEMs at 2 cm (scaling to ~1000 m) and 6 cm (scaling to 3000 m) upfault in the major channels, on the crest lines (Fig. 13A), and at drainage basin head (Fig. 13B). Incision rate in the major channels increases up to a constant value reached after ~70 min at 2 cm upfault and ~90 min at 6 cm upfault, indicating an upstream propagation of the topographic equilibrium (Fig. 13A). Erosion of the adjacent crest lines starts with a delay of ~30 min (scaling to ~0.4 My) at 2 cm upfault and ~120 min (scaling to ~1.6 My) at 6 cm upfault. Hillslope erosion rate rapidly increases to a peak value and then slowly decreases until reaching the same stable value as for the major channels. This observation indicates that the major channels control footwall erosion by imposing the erosion rate.
Table 3
Values of the imposed normal fault slip and time in the experiment for each DEM in the extensional setting experiment; note that the fault dip is 60°.

<table>
<thead>
<tr>
<th>DEM #</th>
<th>Cumulated normal fault slip (mm)</th>
<th>Age (min)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>11.4</td>
<td>30</td>
</tr>
<tr>
<td>2</td>
<td>22.8</td>
<td>60</td>
</tr>
<tr>
<td>3</td>
<td>34.2</td>
<td>90</td>
</tr>
<tr>
<td>4</td>
<td>45.6</td>
<td>120</td>
</tr>
<tr>
<td>5</td>
<td>57.1</td>
<td>150</td>
</tr>
<tr>
<td>6</td>
<td>68.5</td>
<td>180</td>
</tr>
<tr>
<td>7</td>
<td>79.9</td>
<td>210</td>
</tr>
<tr>
<td>8</td>
<td>91.3</td>
<td>240</td>
</tr>
</tbody>
</table>

Hillslope processes respond slowly before reaching the same steady erosion rate as for the major channels with a delay of 90 min (scaling to ~1.2 My). The constant value of erosion rate for the major channels and for the adjacent hillslopes at 6 cm upfault at the end of the experiment (~0.8 cm/h, scaling to ~0.52 mm/yr) leads to the development of a steady state topography (i.e., constant height of triangular facets) that would be preserved if the experiment had continued at the same slip rate. Interestingly, this steady erosion rate represents ~40% of the vertical-component of the imposed fault slip rate (2 cm/h scaling to 1.3 mm/yr). This suggests that the sediment filling close to the fault in the hanging-wall basin is significant and controls the magnitude of erosion in the footwall. During the same period, the evolution of headward erosion increases abruptly up to a peak value around 30 min (scaling to ~0.4 My) with a value of ~19.5 cm/h (scaling to ~12.7 mm/yr) (Fig. 13B).

Then, it decreases following a power law relationship and reaches a stable value of ~4 cm/h (scaling to ~2.6 mm/yr) close to the end of the experiment. Other experiments (Strak et al., 2011) also show that the headward erosion rate is significantly faster than the imposed fault slip rate, indicating a fast upstream migration of the footwall dissection by the channels, while hillslope processes are less efficient and delay the occurrence of a global topographic equilibrium.

3.3. Strike–slip tectonic context

Studies of major active strike–slip faults in Turkey (North Anatolian fault, NAF) or in California (San Andreas fault, SAF) reveal that the fault zone geometry and its morphological evolution control surface fault kinematics as well as its seismic activity (e.g., Barka and Kadinsky-Cade, 1988; Wescoulsky, 1988; Wallace, 1990). We present and discuss hereafter some results obtained from a typical morphotectonic experiment simulating the morphological evolution of a strike–slip fault (Fig. 14 and online material A5 and A6). Such experiments are particularly useful to better constrain how strike–slip fault discontinuities (antithetic faults, fault bends, stepovers) evolve and interact with morphogenetic processes (Chatton et al., 2012). In the following, we focus on the morphologic evolution of the fault zone and surrounding region, the accommodation of horizontal displacements across the shear zone and the control of fault structure on sedimentation to illustrate the potentiality of the experimental approach.

3.3.1. Morphotectonic evolution along a strike–slip fault

The initial topography of the model presented in Fig. 14 was derived from a DEM of the right-lateral Wairarapa strike–slip fault (Pigeon Bush site) in the North Island of New Zealand. Optically stimulated luminescence dating results on this fault yields a dextral-slip rate of ~11 mm/yr and a vertical slip rate of ~1.7 mm/yr at the Pigeon Bush site (Wang, 2001). In the model, fault geometry and kinematics were imposed and then accurately monitored during the whole experiment duration. Fault kinematics was set as pure strike–slip (no dip-slip component), and incremental displacements of 2.8 mm (scaling to ~10 m in nature) were performed every 30 s (scaling to ~300 y in nature) to simulate the cumulated fault offset induced by repeated major earthquakes (M > 7.5).

Surface water runoff generated by the rainfall system induces the development of a drainage network whose geometry is controlled by the preexisting topography of the initial stage, but also by surface deformation because of fault activity and local erosion–sedimentation process interactions. Fault activity induces displacement of topographic features that tend to dam the channels crossing the fault zone (Fig. 15; online material A6). Associated perturbations of the local drainage geometry and surface runoff result into slight but noticeable reorganisation of the watersheds. In the case of a dextral strike–slip fault, channels sourced in high topography enlarge their bed width by preferentially eroding their right bank. As expected, after a certain amount of cumulated fault slip, all the channels crossing the fault zone exhibit right lateral en baionnette geometry.

Surface fault trace is characterised by en-échelon segments at the beginning of the experiment (Fig. 15B), but rapidly evolves toward a more linear and continuous surface rupture (Fig. 15C–E). Under the combined effects of slip accumulation, erosion, and sedimentation, the model surface develops tectonic and morphological structures (online material A6) similar to their natural counterpart, such as Riedel’s shear planes, pressure and shutter ridges, pull-apart basins, alluvial fans and terrace risers. Their space and time evolution can be precisely analysed. Particularly, deformation partitioning, sequential formation of alluvial terraces, stream captures, development of ‘traps’ filling with sediments are observed. New offset markers are created continuously during the experiment. Apparently, however, most of them only record part of

![Fig. 12. Evolution of a major channel-long profile (A, B) and one of its adjacent crest-long profile (C, D). The upper panels used the pristine surface of the footwall as a reference for elevation, while the lower panels used the base level. Note that from DEM 5 (in green), a topographic equilibrium starts to be observed close to the fault.](image)
the local cumulative offset on the fault because erosion and also sedimentation modify their shape. As a consequence, only very rare sites along the fault can be used to retrieve the real cumulated displacements. This discrepancy can be quantified in experiments because the imposed offset and active fault segments are known at each step of model evolution.

3.3.2. Evolution of offset along a strike–slip fault

In compressional and extensional contexts, the fault trace is generally unique and continuous because the fault plane remains relatively localised when it reaches the surface (Figs. 4 and 9). Conversely, surface fault geometry and morphology in strike–slip settings appear more complex because crustal shearing generates a broader deformation.

Fig. 13. Evolution of the erosion rate measured in the major channels and on the adjacent crest lines at 2 and 6 cm upfault (A) and of the headward erosion rate (B). The black arrow indicates the vertical component of the imposed fault slip rate (2 cm/h). Colours are referred to Fig. 9.

Fig. 14. Oblique view of the final stage of the geomorphic experiment in the strike–slip setting. The red box shows the location of Fig. 15. Black arrows indicate fault trace.
zone on both parts of the main fault plane. Several fault branches, that are R and P’ riedel types, and R’ antithetic fault (e.g., Cunningham and Mann, 2007; Fossen, 2010) propagating at a short distance of the main fault interact, generating significant fault slip partitioning. Moreover, recent studies point out a possible correlation between the structural maturity of the fault and the amount of deep slip measurable at the surface (e.g., Dolan and Haravitch, 2014). To illustrate experimentally how deep fault slip is accommodated along a strike–slip fault, we focus hereafter on the kinematics of surface deformation along several fault-perpendicular profiles situated at three different sites along the wrench zone (Fig. 16). Profile A crosses a single fault segment, profile B crosses a releasing bend bounded by two fault branches, and profile C crosses a complex deformation zone where several fault branches join together. In the case where the surface rupture exhibits a well-localised single fault (profile A), measurements of offset markers can lead to a reasonable estimation of the true offset. Indeed, an offset of 2.3 mm can be measured from offset morphologies, which represents 82% of the far field block relative displacement (imposed offset = 2.8 mm). Here, only < 20% of the strike–slip component is therefore accommodated off the fault by diffuse shearing deformation. Along profile B, which is located on a releasing bend bounded by two subparallel faults, the strike–slip component is accommodated by a double fault zone and also by two diffuse shear zones. Each fault accommodates respectively 0.75 mm and 1.0 mm. The first shear zone is located inside the releasing bend whereas the other is located to the north of it. Considering that only sharp shear gradients can be properly measured in the field using morphological markers, offsets along the two fault zones lead to a maximum total amount of 1.75 mm, which represents about 63% of the imposed far field displacement. This underestimation of fault slip is even more critically evidenced when looking at profile C, which is located at a complex junction between several strike–slip branches. Only a gentle sharp gradient of 0.7 mm can be evidenced, meaning that about 75% of the strike–slip component is likely accommodated by diffuse shearing extending over more than 1.5 cm apart from the fault zone (equivalent to a few hundred meters in nature). Experiments highlight that measuring fault offset from morphostructural observations along the fault is not an easy task. As in nature, instantaneous and cumulated fault offsets can be largely underestimated.

3.3.3. Subsurface fault structure

Sectioning the model at the end of the experiment allows for an analysis of the fault geometry at depth and of syntectonic sedimentary layers. This is of interest to better understand the interactions between fault structure and tectonic activity versus surface processes (erosion.
Except in rare areas where the model exhibits a unique fault plane, the strike–slip fault zone is generally composed of several branches that join together at depth into a single vertical plane (Fig. 17). This geometry is similar to the so-called negative and positive flower structures that characterise natural strike–slip fault geometry in seismic profiles (e.g., Harding, 1985; Sylvester, 1988). Depending on the sense of slip and the local orientation of the fault branches, a normal or a reverse component is induced, resulting in the formation of sag ponds and pull-apart basins or pressure and shutter ridges, respectively. As a consequence, bulk displacement accommodated at depth on the basement fault is distributed toward the surface among several faults whose tectonic activity evolves through time. Some branches remain

Fig. 16. Quantifying fault offset, like for instance at a trench site, faces several limitations related to the local complexity of the fault trace and the amount of diffuse deformation accommodated off the main fault. These could result in a significant underestimation of the true offset.
inactive during a certain period, then they are reactivated later when their geometry becomes compatible again with the evolving strain field in the wrench zone. As observed in natural cases, sediment thickness and facies vary significantly on both sides of the main fault branches, which complicate the determination of fault kinematics without combining their analysis with surface morphology. Despite these limitations, some interesting observations arise from the experimental model. For instance, local relief apparently has a clear influence on subsurface fault geometries. In regions of low topography or where strong river incision traverses the wrench zone, the fault trace is deviated toward topographic lows, and a narrow releasing bend generally develops. In regions of higher relief, compression is often associated with shearing, and the fault zone appears much more deformed and segmented. Fault segments are short and interact more intensively. Here, the interpretation of sedimentary records is useless to constrain fault kinematics because, as is the case in the field, only three-dimensional trenching enables the reconstruction of fault slip kinematics.

4. Discussion

4.1. Comparison of models with nature

The aim of experimental modelling in the geosciences is not to reproduce the whole complexity of natural systems but to obtain first-order information on the dynamics of processes and their couplings. Therefore, it is very insightful to point out the similarities and dissimilarities between models and their natural counterparts to assess the relevance of experimental results toward nature (see also discussions in Graveleau et al., 2011; Strak et al., 2011).

Initial model geometry is a key parameter of experiment boundary conditions, which depend on the scientific objectives and also technical limitations. In the strike-slip experiment, a decimetric-high pre-topography was used because the observation time window was focusing exclusively on the last thousand years (Fig. 14A). The total duration of the experiment was not long enough to generate the long-term topographies required to account for the real morphology of the case study. In the compressional and extensional tectonic contexts, the initial stage topography was maintained flat and subhorizontal whereas inherited relief always exists in nature. This solution was adopted to avoid influencing fault nucleation and deformation propagation but also to facilitate parametric studies based on similar and easily reproducible initial stage (Figs. 5B and 10A). In return, a flat subhorizontal topography significantly lengthens experiment duration because such a surface is very difficult to dissect. A long time is needed to acquire enough slope along the active scarps to trigger erosion and activate morphogenesis. As a consequence, remnants of the initial surface are often overrepresented at the end of the experiment.

In most of our models, drainage network morphology slightly differs from the natural one. Major incising channels present generally wider beds (~0.5–5 cm scaling to ~250–2500 m; see Fig. 3) than their natural counterparts (~100–1000 m). This discrepancy already has been highlighted in other geomorphic experiments and notably attributed to the use of water as the sediment transport fluid (Paola et al., 2009). Indeed, using water in models entails low Reynolds numbers compared with nature (e.g., Niemann and Hasibargen, 2005; Graveleau et al., 2011). However, an unscaled Reynolds number is not necessarily problematic as the physics of laminar microscale experimental rivers still account for sediment transport law characterising natural turbulent rivers (Malverti et al., 2008). Another possible element of explanation could be related to the physical properties of our material that present a change in material rheology with depth (e.g., Niemann and Hasibargen, 2005; Graveleau et al., 2011). Indeed, the first top half-centimetre of the model is water saturated and mechanically very weak, whereas a significant downward hardening occurs. The lack of fine particles and vegetation could also be mentioned as causing unsteadiness of the channel banks. Clearly, complementary investigations focused on fluid shear effect in channel beds and its scaling with the shear resistance of the experimental material are still required. One way to reconcile channel width in models with the natural ones could be to consider the average time scaling (1 s ~200 y). Indeed, it appears that experimental channel flow in our model should not be compared with human scale observations of river flow but with major flooding events, which are among the most efficient driving processes of erosion and transport (e.g., Turowski et al., 2009). The wide channels that develop in our experiments should thus be considered as equivalent to floodplains of natural rivers and not as individual rivers. It could be like comparing channel in our models to a photograph of a natural river shot with a secular-long exposure time. At that scale, natural river dynamics would not appear turbulent but likely laminar. Active riverbeds would not be cloistered in the low flow channel but would spread over the whole floodplain.

Furthermore, channel-long geometry in experiments does not always fit natural river trends. For instance, in the extensional experiment, channel-long profiles developing across the normal fault appear convex upward, while river-long profiles in nature are concave upward.
when topography reached a steady-state equilibrium (e.g., Hack, 1957). This difference is still not well understood (see discussion in Strak et al., 2011), but the fact that the topographic equilibrium is not reached over the entire model may play a role. However, the increase in local slope above the thrust hanging-wall (Fig. 6D) is a feature classically observed in nature for similar tectonic setting (Ouchi, 1985). Finally, the shape of the simulated triangular facets is concave upward while it is convex to linear in nature (e.g., Petit et al., 2009; Strak, 2012). This may come from the substantial erosion of the escarpment by mass wasting processes in our model, whereas slower hillslope processes such as soil creep seem to prevail in nature (e.g., Font et al., 2002).

Erosion–transport processes in experiments can be compared to natural ones. The granular composition of the experimental material was determined so that once saturated in water and exposed to water runoff it erodes both by incision owing to channelized runoff and by gravity-driven hillslope processes. In nature, erosion occurs through a wider range of processes depending on topography, climate, vegetation, lithology, land use, etc. (e.g., Carson and Kirkby, 1972). A difference in hillslope dynamics is observed when a topographic gradient of 25° is reached, corresponding to a threshold value where landslide processes are significantly activated (Roering et al., 1999; Montgomery and Brandon, 2002). In our models, we observe two different categories of hillslope processes depending on the slope value. At shallow slopes (<15–20°), diffusive–like processes slowly transfer particles toward the main channels; while at steeper slopes (>15–20°), landside-like mass wasting processes mobilize the top few millimetres of experimental material. The observed slope threshold is correlated to a significant change in mean erosion rate (Graveleau et al., 2011) and recall the general evolution of erosion rate as a function of mean slope in nature (Montgomery and Brandon, 2002). As a consequence, despite this slight discrepancy between nature and model regarding the value of threshold hillslope gradient for activation of landside-like processes, the experimental material succeeds in reproducing the general trend of flux evolution as a function of topographic gradient. Concerning rivers, the mode of transport of solid particles in natural streams is classically divided into two end-members, which are suspended load and bedload (Yalin, 1977). Although suspended load represents a great proportion of drainage basin efflux because of erosion (Allen and Allen, 2005), bedload erosion–transport has been shown as a first-order mechanism of stream incision (e.g., Foley, 1980; Sklar and Dietrich, 2001; Cook et al., 2013). In our model, transport is essentially achieved by bedload and no load is carried in suspension. Therefore, bedload transport in the experiments provides the agent to detach particles and incise along channel beds.

In experiments, transported particles are deposited at different distances from the channel outlet depending on their nature, thereby characterizing different sedimentary environments. Indeed, the experimental material (MatIV) has been designed so that particle grain size, grain shapes and density distribution promotes particle sorting during transport and triggers stratified sedimentation. In addition, the colour difference between material particles allows us to distinguish alluvial features in map view and the bedding, stratigraphic discontinuities, and lateral facies transitions in cross section. As a result, sedimentary deposits display striking details (Figs. 3, 11 and 17). For instance, the fault-perpendicular geometry of the syntectonic deposits in the extensional context exhibits a thickening toward the fault and a progradation of sediments away from the fault, as observed in nature (e.g., Stein et al., 1988; Armijo et al., 1996). Deposits in all models show lobed geometries in map-view that are very similar to natural alluvial fans. Their slope ranges about 3–8°, which is slightly above the range of values of natural alluvial fans in orogenic contexts (1–4°) (Saito and Oguchi, 2005; Graveleau, 2008; Graveleau et al., 2008). Particle segregation during transport provides a proxy of natural sediment grain size. Indeed, silica powder particles (density of 2.65) and glass microbeads (density of 2.5) are deposited in the proximal domain of alluvial fans and represent therefore the coarse-grained sediments (Fig. 3G). Graphite and PVC particles (density of 2.25 and 1.38, respectively) are deposited in the distal domain of fan deposits and can therefore provide a proxy of fine-grained sediments.

Finally, the landforms that developed in our morphotectonic experiments (Fig. 3) show striking similarities with nature, as demonstrated by preliminary morphometric comparisons between experimental features and natural landforms (Graveleau and Dominguez, 2008; Strak et al., 2011). For instance, in the extensional context, the widths of triangular facets (~2.5–8 cm scaling to ~1250–4000 m) are comparable with those observed along natural normal faults, such as the Wasatch fault (Utah) and the Tunka half-graben fault (Western Baikal rift). The topographic profile measured along the major incising channels and along the main crest lines also roughly scale to their natural counterparts (see Fig. 9 of Strak et al., 2011). For experiments developing quasi-linear mountain ranges like in the extensional setting, we notice the morphological similarity of the contiguous major watersheds and also their regular spacing. This is favoured by the lack of preexisting topography and determined during the early stages of channel network growth (Hovius, 1996; Castelltort and Simpson, 2006). A detailed morphometric analysis and comparison of the experimental and natural landforms would be certainly very insightful. Some early tests already have been carried out regarding for instance, Hack’s law (Hack, 1957) or the mountain front sinuosity (Bull and McFadden, 1977). For the former, a similar power law relationship is observed in nature and in models with equivalent Hack exponents close to 0.5 (see Fig. 7 of Strak et al., 2011). For the latter, the sinuosity of the fault trace (Smf) is close to 1 in the extensional setting and slightly larger in the compressional setting (Smf ~ 1.2), meaning that both piedmonts are tectonically active and experienced sustained uplift. Further morphometric investigations are still needed, notably to investigate how the drainage density (Horton, 1945), the hypsometry (Strahler, 1952), and the ratio of valley floor width to valley height (Bull and McFadden, 1977) evolve through time.

4.2. Morphotectonic marker formation, evolution, and record of deformation

One of the major valuable outcomes of the experimental approach is to provide insights on how morphotectonic markers progressively form, evolve, interact, and record deformation. In the compressional setting, morphotectonic models simulate successfully the evolution of terrace forming above an active thrust (Figs. 3I and 5). This remarkable graphical sequence is consistent with the classical model of fluvial terrace nucleation and progressive uplift above an active thrust or fault-related fold (e.g., Avouac and Peltzer, 1993; Avouac et al., 1993; Lavé and Avouac, 2000; Daéron et al., 2007; Simoes et al., 2007; Li et al., 2013). Models allow us to observe how progressive channel incision and continuous uplift of the hanging-wall trigger the abandonment of the channel bed and the formation of terraces. This dynamic is associated with avulsion on alluvial deposits and channel piracy likely associated with changes in flow stream power. Because of the steady uplift of the hanging-wall, preserved terraces remain parallel to the active riverbed (Fig. 6), a feature not always observed in the field (Le Béon et al., 2014; Simoes et al., 2014). Finally, uplift rate inferred from terrace geometry is in perfect agreement with the uplift rate determined using the horizontal shortening rate and the geometry of the thrust. This methodology is classically used in the field to infer fold growth and fault slip rates (e.g., Molnar et al., 1994; Poisson and Avouac, 2004).

In the extensional setting model, the progressive mergence of triangular facets and watersheds during the early stage of topographic evolution has also been observed in numerical models (Petit et al., 2009). In nature, the triangular facets form and evolve by slope retreat (Hambly, 1976). Steep slope forming at the base of the facet progressively decreases toward the facet summit, leading to a convex upward profile. In the model, however, the steep slope formed at the base of the facet is rapidly eroded by mass wasting processes, producing a concave upward profile. Furthermore, the rapid development of a steady
number of triangular facets and watersheds with constant hillslope gradients close to the fault is striking and suggests that normal natural fault scarp could potentially display rapidly steady along-strike topographic features, even if the further uplift topography is still in a transient regime. Finally, a relationship has been evidenced in the model between the height of triangular facets and the fault slip rate. This has been also evidenced in the field (Menges, 1990) and in numerical models (Petit et al., 2009).

In the strike–slip setting, the formation and evolution of specific morphologic markers (such as pressure ridges and offset drainage channels) appear to be well simulated considering the good analogy between simulated and natural morphologies. The observed progressive diversion of the drainage network crossing the dextral strike–slip fault illustrates the classical model of beheaded stream formation and evolution (Sieh and Jahns, 1984; Grapes and Wellman, 1988).

Finally, the compressional and extensional tectonic contexts show interesting features of geomorphic evolution over a pristine subhorizontal surface. Formation and growth of the drainage network is controlled by headward migration of watersheds, suggesting the occurrence of a similar process in nature when an old tabular erosional surface is progressively uplifted and dissected (e.g., Kennan et al., 1997; Coltorti and Ollier, 1999; Vassallo et al., 2007).

4.3. Landscape response time scales

Time scales of landscape responses to tectonic forcing have been evaluated for models conducted at short time scales (i.e. earthquake in the strike–slip setting) and long time scales (i.e. piedmont evolution in the compressional and extensional settings). A different time scaling, that is 1 s – 50 y in the former and 1 s – 200 y in the latter, is proposed. A difference for spatial scaling is also considered, which is 1 cm – 50 m at the earthquake spatial scale and 1 cm – 500 m at the scale of a mountain front. Time and spatial scaling in the experiments can be adjusted depending on the temporal and spatial scales of the studied object. These scalings are rough estimations that have to be considered with caution. However, these chosen time scales allow delivering averaged deformation, erosion, and sedimentation rates in the range of natural equivalents, indicating that a comparison between model and nature is relevant. Particularly, we stress that our approach provides a way to observe some temporal relationships between deformation and surface processes that would be difficult to observe in the field. This concerns for instance (i) the typical lifetime of morphotectonic markers, (ii) the regularity of marker formation in response to steady tectonic forcing, and (iii) the evolution in space and time of erosion rates.

First, one of the common observations made in the three tectonic contexts is that some morphological markers (such as terraces, sag ponds) can be very ephemeral compared to the total duration of the experiment. This is observed in the field, but the average preservation time of such morphotectonic markers is difficult to establish. For instance, some terraces in the compressional context had a lifetime of a few seconds to tens of minutes between their formation and removal by erosion (equivalent to 1–500 ky in nature). Features that were preserved at the end of the experiment (or to the present day, in nature) only provide a very fragmentary sampling of the morphotectonic marker population that actually formed along the active structures.

Second, another important observation made in the compressional setting experiment is that terraces form irregularly along channels despite tectonic and climatic forcings that were held constant. This unsteadiness of the geomorphic system that emerges from a steady tectonic forcing is observed in the field, but the scales and rates remain poorly known (e.g., Sweeney et al., 2012). Further experiments focused on the dynamics of terrace formation (e.g., Mizutani, 1998; Hancock and Anderson, 2002) and particularly the factors controlling the autogenic or allogenicity of this marker in all tectonic contexts would be very insightful to constrain the scales and rates of such unsteady record. Complementary investigations could for instance explore the reproducibility of this unsteadiness with multiple model runs using the same setup, parameters, and boundary conditions. This would be a measure of chaotic or nonlinear processes leading to the unsteadiness.

Third, measuring erosion rates in the field over a wide range of spatial and temporal scales is a great matter of research challenge to distinguish the steady or transient states of topography evolution (e.g., Kirchner et al., 2001; Koppes and Montgomery, 2009; Cook et al., 2014). The regular monitoring of topography throughout the experiments allows us to investigate both the spatial and temporal evolution of erosion and sedimentation rates. Our approach could therefore provide a way to compare instantaneous and long-term averaged rates and to document the transient and equilibrium phases of topography evolution. In the example of a punctual quantification of erosion through time in the extensional context, we can help resolve two questions: (i) “What is the dynamics of landscape evolution from transient state to steady-state?” and (ii) “What is the ratio of erosion/tectonic fluxes in the long term?” First, we have shown that a topographic equilibrium is reached first in the hanging-wall basin and progressively propagates upstream, characterising the end of the growth phase of the relief that tends toward a steady state. The portion of relief where a topographic equilibrium is observed evolves still locally through short-term surface processes such as landslides and channel avulsion. Furthermore, we have evidenced in our experiments that the erosion rate measured in the long term is 0.4 times the imposed vertical fault slip rate, highlighting the role of sedimentation on erosion upfault (Babault et al., 2005).

4.4. Possible improvements and lines of further research

Modelling of morphotectonic processes always evolves. Depending on targeted scientific issues and modelling outcomes, improvements of experimental setups and procedures can be achieved. For instance, although we consider that the present compositions of the experimental material give good results in terms of similarities of processes and landforms between nature and models (i.e., Graveleau et al., 2011; Strak, 2012), these compositions and associated physical properties can still be improved. For instance, it would be good to add in the analogue material other particles with densities between plastic (1.38) and graphite (2.25). Similarly, modulating silica powder and glass microbead grain sizes could help reduce the detachment threshold for incision that is presently too high. These would enhance erosion of low slopes by decreasing the average detachment threshold for incision and would increase the average transport distance of sedimentary deposits. It could also help preserve slopes steeper than 20° and thus better simulate hill-slope gradients observed in nature. In all cases, improvements in material composition could also help develop the investigation of sedimentary basin stratigraphy and notably their recording of the evolution of surrounding topography. We have not deeply investigated this aspect in our modelling, but the approach has been used already to link watershed fluxes to the geometry of alluvial fans or deltas (Heller et al., 2001; Paola et al., 2001; Raynal et al., 2007; Rohais et al., 2012). Integration of sequence stratigraphy analysis in our modelling approach would surely be insightful in a source-to-sink model integrating the interactions between tectonics, topography, erosion, and sedimentation.

Water is a very useful liquid for saturating the experimental material and for eroding the model. However, some scaling issues already mentioned above suggest that pure water is actually not an ideal liquid for small-scale experimental modelling. A working possibility would be to test the addition of adjuvant that would accordingly modify liquid viscosity or surface tension forces. This could decrease the capillarity forces within the experimental material composing the topography of the model, which could also contribute to increase erosion rates and modify the shape of the model landforms.

Finally, further lines of research that we partly explored in this paper can be greatly expanded with the continuous recording of the kinematic
and topographic evolution of the models. For instance, a continuous knowledge of the horizontal and vertical displacement fields allows us to investigate the spatio-temporal variability of the accommodation of deformation along major faults and their expression in the topography. Scientific questions addressing the influence of surface processes and topographic evolution on the partitioning of deformation along each type of faults could be explored. Similarly, continuous monitoring of the topography evolution could provide a measure of the evolution of fluxes in a source-to-sink approach, from the drainage divide to the sedimentary basin. This would bring significant insights on the mechanisms of solid load transfer between zones undergoing erosion and transient or steady storage.

5. Conclusion

Experimental modelling allows studying at different time and spatial scales the impact of tectonics, erosion, and sedimentation interactions on landscape evolution. In this paper, we have investigated the formation and evolution of morphotectonic markers that develop in different tectonic contexts (thrusting, normal faulting, and strike–slip faulting) to illustrate the potentialities of this approach. In the compression setting, our experiment reveals that the formation of stream terraces is controlled by hanging-wall uplift, channel incision, and avulsion on the alluvial fans. Similar patterns of terrace deformation above active thrusts or fault-related folds between the experiment and the field are observed. However, terraces appeared irregularly through the experiment duration despite tectonic rates and erosional conditions being held stable. Complementary investigations are needed to constrain how the unsteadiness of processes in the geomorphic system can emerge from the steadiness of tectonic forcing. In the extensional setting, sedimentary filling in the hanging-wall controls football erosion and triangular facet development through the position of the base level and the development of topographic steady-state conditions. To illustrate the evolution of drainage network along an active right-lateral wrench fault zone. They shed light on methodological issues, such as uncertainty assessment when calculating slip rates from offset markers. They reveal that a systematic underestimation of several tens of percent should be likely considered depending on the local geometry of the surface rupture and the characteristics of surface deformation.

Our results indicate also that the experimental approach provides valuable insights about the morphological dynamics of topography evolution for all tectonic settings. It can be used to constrain the mechanisms of morphogenesis that are difficult to assess when observing a landscape in the field or on any aerial images. Relief dynamics is particularly illustrated qualitatively by movies of the experiments and quantitatively by DEIs and kinematic analyses. Such a dynamic view could be very useful and enlightening for a wide range of geoscientists and a broad public involved in environmental and land use planning. Geomorphic experiments could provide a guide to neotectonic, palaeo- and geomorphological studies to help fill the gap between short-term and long-term observations of tectonic deformations and relief growth.

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