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Slip acceleration on normal faults due to erosion and sedimentation — Results from a new three-dimensional numerical model coupling tectonics and landscape evolution

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ABSTRACT

Although it is well accepted that mass redistribution on Earth's surface due to erosion and sedimentation affects the rate of crustal deformation, the influence of surface processes on the slip behaviour of individual faults remains poorly understood. Here we use the new tool CASQUS, which combines landscape evolution modelling with three-dimensional tectonic models, to investigate how erosion and sediment deposition affect fault behaviour in extensional tectonic regimes. A systematic set of experiments with models consisting of an isostatically responding elastic upper crust containing a single normal fault shows that the slip accumulation on the fault is accelerated by erosion and sedimentation. Depending on fault geometry, extension rate and parameters controlling erosion and sediment deposition, the slip rate of the normal fault is up to ~15% higher when surface processes are taken into account. The slip acceleration can be explained by an increase in the differential stress caused by erosion of the fault footwall and sedimentation in the hanging wall, which alter the faulting-induced flexure of the crust. We observe a similar slip rate increase in a more complex model of a growing horst bounded by three faults of different length. Our results show that a positive feedback exists between surface processes and tectonics and imply that geologically derived fault slip rates that integrate over several earthquake cycles may contain a contribution from surface processes.

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1. Introduction

Slip accumulation on normal and thrust faults at depth is commonly expressed by the growth of topography on Earth's surface (King *et al.*, 1988; Stein *et al.*, 1988; Anders *et al.*, 1993; Densmore *et al.*, 2004; Hetzel *et al.*, 2004; Ellis and Densmore, 2006). The resulting topography may range from fault scarps generated by single earthquakes over fault-bounded mountain ranges to large mountain belts involving many faults. Regardless of the spatial dimension, the fault-created topography will ultimately be the target of erosion and sedimentation processes, which will lead to a redistribution of mass at the surface. The mass redistribution may, in turn, considerably influence the localization and style of crustal deformation, as demonstrated by numerical modelling on different scales (Koons, 1989; Beaumont *et al.*, 1992; Masek *et al.*, 1994; Koons, 1995; Avouac and Burov, 1996; Densmore *et al.*, 1998; Batt and Braun, 1999; Willett, 1999; Garcia-Castellanos, 2002; Fischer *et al.*, 2004; Simpson, 2004a,b; Pysklywec, 2006; Burov and Toussaint, 2007; Braun *et al.*, 2008).

Although these numerical models provided important insights into the interaction between crustal deformation and surface processes, coupling between erosion, sedimentation and

tectonics with an adequate representation of these processes in three dimensions has not yet fully been realized. Instead, models investigating the feedback mechanism between tectonics and surface processes can be roughly divided in two main types: (a) Tectonic, typically orogenic-scale models, which are two-dimensional and apply one-dimensional erosion laws to the topographic profile of the model orogen (e.g. Beaumont *et al.*, 1992; Willett, 1999; Pysklywec, 2006; Tomkin, 2007) and (b) landscape evolution models, which may include surface processes in two dimensions but commonly implement tectonics by simple, prescribed displacement functions (e.g. Kooi and Beaumont, 1996; Tucker and Slingerland, 1996; Densmore *et al.*, 1998; Simpson, 2004a,b). With respect to the question how faulting is affected by erosion and sedimentation, neither type of model can make quantitative predictions because faults are not included in these models. As a consequence, the impact of surface processes on the slip evolution of individual faults remains an open issue.

Here we present a new numerical modelling approach, which fully couples in three dimensions a landscape evolution model and a tectonic model that may include one or more discrete faults. The intention of our study is twofold: First, we introduce the basic principles and capabilities of our approach that opens new horizons for investigating the interaction between tectonics and surface processes. Second, we use the new tool to systematically study the effect of erosion and sediment deposition on the slip evolution of a single normal fault. Finally, we present an example of a more complex tectonic setting with three normal

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faults of different length forming a horst to show how the slip accumulation controls topography and stream directions and how individual faults respond to the surface processes.

2. Modelling principles and model setup

2.1. Principles of CASQUS

The new software tool called CASQUS allows to investigate the interaction between tectonic and surface processes in three dimensions by integrating the erosion and sediment transport routines of the landscape evolution model CASCADE (Braun and Sambridge, 1997) into geomechanical models created with the commercial finite-element software ABAQUS (Hibbitt et al., 2008). In the following we will summarize the principles of CASQUS; technical details concerning the implementation of CASCADE into ABAQUS are provided by Kurfeß (2008) and Kurfeß and Heidbach (Computers and Geosciences, in press). The basis for CASQUS is a three-dimensional ABAQUS model, which represents the tectonic setting and may include gravity, isostasy, different rheological layers as well as faults (e.g. Hampel et al., 2007; Maniatis and Hampel, 2008). Owing to the ongoing tectonic deformation, the elevation of the ABAQUS model surface will change during a model run. The current elevation of the model surface

serves as input for the landscape evolution model, which calculates the amount of erosion or sedimentation at each point. The information on the change in elevation resulting from the surface processes is transferred back into the ABAQUS model to shift the surface nodes to their new elevation. Shifting of the surface nodes is performed using an Arbitrary Lagrangian–Eulerian approach and adaptive meshing for the associated elements in order to prevent arbitrary stress changes. Hence the shift of the surface nodes causes only changes in the volume and thus in the mass of the associated elements. These mass changes, in turn, cause an isostatic response of the model and alter its stress field, which affects the ongoing tectonic processes. As the latter create the topography that is the target of the surface processes a truly bidirectional feedback exists between surface and tectonic processes in the CASQUS models.

The landscape evolution model used by CASQUS includes diffusive hill-slope processes and fluvial erosion and deposition. The former is implemented by the linear diffusion equation, in which the temporal change in elevation of a point is proportional to the second spatial derivative of topography (Kooi and Beaumont, 1994; Braun and Sambridge, 1997):

$$\frac{\partial h}{\partial t} = k_D \nabla^2 h \quad (1)$$

Fig. 1. a) Setup of the three-dimensional CASQUS model with an isolated normal fault that is embedded in an elastic upper crust and cuts the model surface. Slip initiation on the fault is controlled by the Mohr–Coulomb criterion and caused by extending the model at a rate of $v=2$ mm/a. Isostasy is implemented by a lithostatic pressure and an elastic foundation at the bottom of the model. The bottom of the model is free to move in the vertical and the horizontal directions; model sides in the xz -plane are fixed in the y -direction. Grey layer marks the uppermost layer of elements, which is affected by erosion or sedimentation at the model surface as calculated by the landscape evolution model. Symbols are ρ density, E Young's modulus, μ coefficient of friction, g acceleration due to gravity and P lithostatic pressure. **b)** Mesh of the finite-element model. Shaded in grey is the top element layer that is affected by the surface processes.

Fig. 2. Perspective view of the model surface after 750 ka of extension at a rate of 2 mm/a **a)** without surface processes being active and **b–d)** with surface processes. The diffusion constant k_D in the experiments with surface processes is **b)** 0.1 m²/a, **c)** 0.3 m²/a and **d)** 3 m²/a, respectively. The parameter that controls the fluvial erosion is set to a constant value of 0.03 m/a. Note that with increasing diffusion constant, the resulting topography becomes smoother. Vertical exaggeration is 5.

where h is elevation, t time and k_D the diffusion constant, which depends on climate conditions and lithology. Eq. (1) describes the cumulative effect of different types of surface processes such as slope wash, rain splash and soil creep.

The temporal change in elevation at a point on the model surface caused by fluvial erosion, transport and deposition is calculated by the following relationship

$$\frac{\partial h}{\partial t} = \frac{1}{l_f} (q_{f,eqb} - q_f) \quad (2)$$

where q_f is the sediment flux of a river, $q_{f,eqb}$ the sediment transport capacity and l_f the erosion/deposition length scale. The latter is an empirical constant that describes how easily the river substrate can be eroded or how fast sediments are deposited in a river and thus affects how fast the river locally tends towards equilibrium. Erosion occurs for $q_f < q_{f,eqb}$ and deposition for $q_f > q_{f,eqb}$. The fluvial model is transport-limited, i.e. q_f cannot exceed $q_{f,eqb}$ (Braun and Sambridge, 1997). The sediment transport capacity $q_{f,eqb}$ is calculated by

$$q_{f,eqb} = -K_f q_r \frac{dh}{dl} \quad (3)$$

where K_f is a non-dimensional empirical transport coefficient and q_r the fluvial discharge, which depends on the catchment area and the precipitation rate p , and dh/dl the slope in the direction of river drainage. The parameter that ultimately controls fluvial erosion in the model enters CASQUS as the product of the transport coefficient K_f and the precipitation rate p . An irregular mesh is used at the model surface to avoid the directional bias, which arises from the development of the stream network along preferred directions (cf. Braun and Sambridge, 1997). Sediments may leave the model via the model boundaries.

It should be noted that the erosion–deposition formulation used in CASCADE by Braun and Sambridge (1997) and initially introduced by Kooi and Beaumont (1994) calculates the first-order pattern of erosion and sediment deposition. In other words,

it takes into account the capacity of the river to erode and transport sediment but not the small-scale physics of bedrock erosion.

2.2. Setup of the three-dimensional model with a normal fault

Fig. 1 shows the setup of the model including a normal fault, which we used in our present study. The model represents a 15-km-thick elastic upper crust, which is 100 km by 100 km wide (Fig. 1a). The uppermost 2 km of the model consist of a layer of hexahedral elements, which have an edge length of ~ 1 km at the model surface and define the irregular grid required by the landscape evolution model (Fig. 1b). The remaining part of the model is meshed with tetrahedral elements with decreasing mesh density toward the bottom. Gravity is applied as a body force and the bottom of the model is allowed to move in the vertical and horizontal directions. Isostatic effects are included by applying a lithostatic pressure and an elastic foundation to the bottom of the model. The lithostatic pressure counterbalances the weight of the overlying upper crust whereas the elastic foundation simulates the density contrast between the upper and lower crust, which generates buoyancy forces in response to deformation of the upper crust. In the centre of the model, a planar fault, which cuts the entire upper crust and the model surface, is included (Fig. 1a). The length and the dip of the fault are varied in different experiments. Slip initiation on the fault is controlled by the Mohr–Coulomb failure criterion

$$\tau = c + \mu \sigma_n \quad (4)$$

where τ is the shear stress on the fault plane, c cohesion, μ friction coefficient and σ_n the normal stress on the fault plane. In our experiments, the friction coefficient is $\mu = 0.4$ and the cohesion is zero because fault rocks have negligible cohesion (e.g. Twiss and Moores, 1992). The definition of a maximum shear stress on the fault through the Mohr–Coulomb criterion implies that the stress level on the fault is lower than in the surrounding crust, which is elastic and hence has infinite strength. In order to establish an extensional tectonic regime, the model is horizon-

Fig. 3. Slip evolution at the centre of the model normal fault at 2 km depth without and with surface processes. Experiments with surface processes were calculated with diffusion constants of 0.1, 0.3 and 3 m²/a and a fluvial erosion parameter of 0.03 m/a. The length of the 60°-dipping fault is 40 km. Note that the fault slips faster if erosion and sedimentation are active on the model surface.

Table 1

Results from experiments with an isolated normal fault of different length.

Fault length (km)	Extension rate (mm/a)	Slip rate at steady-state (mm/a)			
		Slip rate increase (in mm/a and in % relative to the rate without surface processes)			
		No surface processes	Diffusion constant (m^2/a)		
			0.1	0.3	3
10	2	0.259	0.286+0.027 (10.4%)	0.286+0.027 (10.4%)	0.286+0.027 (10.4%)
	1	0.137	0.152+0.015 (10.9%)	0.152+0.015 (10.9%)	0.152+0.015 (10.9%)
20	2	0.490	0.533+0.043 (8.8%)	0.555+0.065 (13.3%)	0.555+0.065 (13.3%)
	1	0.269	0.307+0.038 (14.1%)	0.307+0.038 (14.1%)	0.307+0.038 (14.1%)
40	2	0.739	0.793+0.054 (7.3%)	0.829+0.090 (12.2%)	0.837+0.098 (13.3%)
	1	0.406	0.454+0.048 (11.8%)	0.465+0.059 (14.5%)	0.465+0.059 (14.5%)

Slip rates are values at steady-state, i.e. after 250 ka of model time, and given with three decimal digits to show the differences between the experiments.

tally extended at a constant rate of 2 mm/a (Fig. 1a). The applied gravity and lateral extension lead to a vertically oriented maximum principal stress σ_1 and a horizontally oriented minimum principal stress σ_3 that initiate normal slip on the fault. The resulting footwall uplift and hanging-wall subsidence generates the topography that is subjected to the surface processes. Note that the faults are not allowed to propagate laterally or at depth during an experiment, i.e. their boundaries are fixed. This approach follows the growth model proposed by Walsh et al. (2002) who suggest that the length of a fault is nearly constant from an early stage, with the consequence that growth is achieved mainly by an increase in displacement. To ensure that the ratio between fault displacement and length does not exceed a realistic value as given by fault scaling laws (e.g. Cowie and Scholz, 1992a,b,c; Schlische et al., 1996), we limit the model time of the experiments to 750 ka. Hence our present models apply to young faults and youthful landscapes and thermomechanical effects, which may become important on timescales of millions of years, are not considered.

3. Model results

In the following we will present the results from a set of experiments, in which we varied the diffusion constant, the fluvial erosion parameter, the fault length, the extension rate, and the

fault dip. The alluvial and bedrock erosion length scales are constant in all experiments and set to values of $l_{f,a} = 10$ km and $l_{f,b} = 100$ km, respectively (cf. Braun and Sambridge, 1997; van der Beek and Braun, 1998). Each experiment starts with the establishment of isostatic equilibrium and then lasts 750,000 model years (750 ka) during which the model is horizontally extended. Surface processes are activated after 100 ka of model time. We performed experiments without and with surface processes to compare the resulting slip histories of the model fault. In all experiments, the fault starts to slip and reaches a steady-state slip rate within 250 ka after the onset of extension. Note that the amount of slip and the slip rate are not prescribed but develop freely in response to the body forces and the velocity boundary condition applied to the model sides and the elastic properties of the crust. Owing to the latter, the total elastic strain resulting from the imposed velocity boundary conditions is taken up as elastic strain by the crust and as slip by the fault. To evaluate the slip evolution of the fault in the presence or absence of surface processes, we extracted the slip history of the fault at a depth of 2 km, i.e. below the layer of elements that are affected by surface processes, and calculated the steady-state slip rate between 250 and 750 ka. Note that we will report the slip rates with three decimal digits to reveal even subtle differences between the experiments.

Fig. 4. Slip evolution at the centre of the fault at 2 km depth extracted from experiments, in which the product of fluvial transport coefficient K_f and precipitation rate p was varied between 0.003, 0.03 and 0.3 m/a; without and with surface processes. The length of the 60°-dipping fault is 40 km. The diffusion constant is $k_D = 0.3 \text{ m}^2/\text{a}$.

3.1. Influence of the diffusion constant

The first model set comprises experiments with a 40-km-long, 60°-dipping fault, in which the diffusion constant k_D was varied between 0.1, 0.3 and 3 m²/a; the parameter controlling fluvial erosion is kept constant with a value of $K_f \cdot p = 0.03$ m/a (cf. Braun and Sambridge, 1997; van der Beek and Braun, 1998). Fig. 2 shows the topography of the model surface after 750 ka. In the absence of surface processes, the fault footwall is uplifted by up to 218 m and tilted away from the model fault, whereas the hanging wall has subsided by 228 m (Fig. 2a). If surface processes are active, this fault-generated topography is modified by erosion and sediment deposition (Fig. 2b–d). The foreflank of the fault footwall is cut by rivers that directly flow toward the hanging wall, whereas the rear flank is cut by a network of rivers, which flow toward the model boundaries or around the fault tips toward the hanging wall. As a consequence, the hanging wall receives sediments from the fault scarp and from the streams flowing around

the fault tips. Additional sediment is delivered from distant hanging-wall regions. With increasing diffusion constant, the topography becomes smoother (Fig. 2b–d). Analysis of the slip evolution shows that the fault slips at a steady-state rate of 0.739 mm/a if surface processes are absent (Fig. 3). In contrast, slip is accelerated if erosion and deposition are active. The slip rate increases relative to the steady-state rate without surface processes by 7.3%, 12.2% and 13.3% for diffusion constants of 0.1, 0.3 and 3 m²/a, respectively, i.e. slip acceleration becomes more pronounced with an increasing diffusion constant (Fig. 3, Table 1).

3.2. Influence of the parameter controlling fluvial erosion

In a set of three experiments with a 40-km-long, 60°-dipping fault, we varied the parameter that controls fluvial erosion in the model, i.e. the product of the transport coefficient K_f and the precipitation rate p , to investigate its effect on the evolution of

Fig. 5. a) Results from experiments, in which the length of the 60°-dipping fault was varied between 10, 20 and 40 km. The slip evolution is shown for the fault centre at a depth of 2 km. The diffusion constant is $k_D = 0.3$ m²/a and the fluvial erosion parameter $K_f \cdot p = 0.03$ m/a. **b)** Results from the experiments with extension rates of 1 and 2 mm/a. The slip evolution is shown for the centre of the 40-km-long, 60°-dipping fault at a depth of 2 km. The diffusion constant is $k_D = 0.3$ m²/a and the fluvial erosion parameter $K_f \cdot p = 0.03$ m/a.

Fig. 6. Results from experiments, in which the dip of the 40-km-long fault was varied between 45°, 60° and 75°. The slip evolution is shown for the fault centre at a depth of 2 km. The diffusion constant is $k_D = 0.3 \text{ m}^2/\text{a}$ and the fluvial erosion parameter $K_f \cdot p = 0.03 \text{ m/a}$.

the model fault.

We used values of 0.003, 0.03 and 0.3 m/a, i.e. we decreased and increased the value of 0.03 m/a used by *Braun and Sambridge* (1997) by one order of magnitude, respectively. The diffusion coefficient used was $k_D = 0.3 \text{ m}^2/\text{a}$ in all three experiments; the extension rate was 2 mm/a. For a value of $K_f \cdot p = 0.3 \text{ m/a}$ the slip rate (0.820 mm/a) is 11.0% higher than without surface processes (Fig. 4). If $K_f \cdot p$ is set to values 0.03 m/a or 0.003 m/a the slip rate is 0.829 mm/a and 0.841 mm/a, respectively, and thus 12.2% and 13.8% higher than the slip rate in the absence of surface processes.

3.3. Effect of fault length

To evaluate the influence of fault length, we carried out additional experiments with fault lengths of 10 and 20 km, using the same fluvial erosion parameter (0.03 m/a) and diffusion constant as in the first model set with the 40-km-long fault (0.1, 0.3 and 3 m/a). As an example, Fig. 5a shows the slip evolution on the 10-, 20- and 40-km-long faults for a diffusion constant of 0.3 m/a. The comparison of all experiments (Table 1) shows that, in the absence of surface processes, the 10-km-long and 20-km-long faults slip at 0.259 and 0.490 mm/a, i.e. at lower rates than the 40-km-long fault (Section 3.1). Erosion and sedimentation increase the slip rate on the 10-km-long fault by 10.4%, regardless of the diffusion constant (Table 1). Slip on the 20-km-long fault is accelerated by 8.8% for a diffusion constant of 0.1 m/a and by 13.3% for diffusion constants of 0.3 and 3 m/a.

3.4. Effect of extension rate

In the fourth set of experiments, we evaluated how the extension rate influences the response of the fault to erosion and sedimentation. Again, we used faults with lengths of 10, 20 and 40 km

and applied diffusion constants of 0.1, 0.3 and 3 m/a; the fluvial erosion parameter is 0.03 m/a. The results show that owing to the lower extension rate of 1 mm/a the steady-state rate on all faults is lower than in the model with an extension rate of 2 mm/a (Table 1; Fig. 5b). However, the percentage of the slip rate increase caused by the surface processes is consistently higher than in the experiment with a higher extension rate. Under the influence of erosion and sedimentation, the slip rate on the 10-km-long and 20-km-long faults increases by 10.9% and 14.1%. Slip on the 40-km-long fault is accelerated by 11.8% for a diffusion constant of 0.1 m/a and by 14.5% for diffusion constants of 0.3 and 3 m/a.

3.5. Influence of fault dip

In the last set of experiments we varied the dip of the 40-km-long fault between 45°, 60° and 75° while applying a horizontal extension rate of 2 mm/a, a diffusion constant of 0.3 m/a and a value of 0.03 m/a for the product of the fluvial transport coefficient and precipitation rate. Inspection of the fault slip evolution shows that the 45°-dipping fault slips at the highest slip rate of all three faults. However, the slip rate increase induced by erosion and sedimentation is highest (12.2%) for a 60°-dipping fault (Fig. 6). The faults dipping at 45° and 75° increase their slip rates by 7.4% and 4.2% respectively, if surface processes are present.

4. Discussion

By combining three-dimensional tectonic models including discrete faults with landscape evolution models, our new modelling technique provides, for the first time, insight into the effect of geomorphic processes on fault slip rates. Previous studies investigating the effect of surface processes on the deformation of the solid Earth showed that the stress and strain distribution is con-

Fig. 7. Plots illustrating the non-linear dependence of the slip rate increase on a) the diffusion constant k_D and b) the product of transport coefficient and precipitation rate $K_f \cdot p$. The slip rate increase is given in percent with respect to the experiment without surface processes.

siderably modified by erosion and deposition, with the consequence that deformation may be locally promoted or suppressed (Beaumont *et al.*, 1992; Densmore *et al.*, 1998; Willett, 1999; Koons *et al.*, 2002; Buiter and Pfiffner, 2003; Simpson, 2004a,b; Pysklywec, 2006; Tomkin, 2007). However, predictions on the fault response to surface processes were not possible because the previous models either included the tectonic motion as prescribed displacement or approximated the faults as shear zones. Our models do not only allow the quantification of the fault behaviour in terms of displacements and slip rates but also reveal that erosion and deposition have a considerable impact on faulting also in extensional regimes.

4.1. Feedback between faulting and surface processes

The numerical experiments with an isolated normal fault show an up to ~15% increase in the slip rate in response to erosion and sedimentation. Although most present methods may not allow the measurement of fault slip rates beyond a 10% uncertainty, the increase in slip rate by erosion and sedimentation is significant. For example, a 20-km-long fault that slips at 0.5 mm/a would accumulate – for a slip rate increase of 15% – an additional displacement 0.75 m in 10 ka, which is equivalent to an additional earthquake of magnitude ~6. Note that owing to the continuous

slip of the fault, our model is limited to settings, in which the earthquake recurrence interval is short with respect to the time period over which the slip rate and surface processes are integrated. In other words, single events with large displacements and the induced erosional impulse on the footwall are not captured by the model.

The magnitude of the slip rate increase is controlled by the fault length and dip as well as by the extension rate, the diffusion constant and the product of transport coefficient and precipitation rate. Shorter faults respond with a relatively high slip rate increase (~10%) already for a diffusion constant of 0.1 m²/a but their slip rate does not increase further with higher diffusion constants (Table 1). In contrast, longer faults exhibit a lower slip rate increase than shorter faults for a diffusion constant of 0.1 m²/a but the slip rate increase relative to the steady-state rate is more pronounced for higher diffusion constants. The dependence of the slip rate increase on both the diffusion constant and the parameter controlling fluvial erosion is non-linear, as the example of the 40-km-long fault shows (Fig. 7). A higher diffusion constant amplifies the slip rate increase (Fig. 7a), whereas a higher fluvial erosion parameter $K_f \cdot p$ causes the slip rate increase to be less pronounced (Fig. 7b). The reason for this contrasting effect is that for a higher diffusion constant erosion and sedimentation are intensified near the fault, whereas an increase of the fluvial erosion parameter enhances the material transport

Fig. 8. Topographic profiles and vertical displacement rate derived from the experiments with a 40-km-long, 60°-dipping normal fault after 750 ka (cf. Figs. 2, 3). a) Topographic profiles across the model centre extracted from the experiments with and without surface processes. Experiments with surface processes used a diffusion constant of 0.1, 0.3 and 3 m²/a, respectively, and a constant fluvial erosion parameter of 0.3 m/a. The profiles show the topography at the model surface and at the top of the tetrahedral element layer (Fig. 1b), respectively; the location of the profiles is indicated in the inset. b) Cross-sections of the model through the centre of the fault showing the vertical displacement rate without (upper row) and with surface processes (lower row; $k_D = 0.3 \text{ m}^2/\text{a}$; $K_f \cdot p = 0.03 \text{ m/a}$).

away from the fault and out of the model. The fault is most sensitive to surface processes if it dips at 60°, which is the optimal orientation of failure with respect to the principal stresses and the typical fault dip for normal faults in nature. In case of a shallower or steeper dip, the response of the fault to the surface processes is less pronounced. Finally, a comparison of the experiments with horizontal extension rates of 1 and 2 mm/a shows that erosion and deposition have a stronger effect on faulting in case of lower deformation rates (Table 1).

To evaluate the sensitivity of our model results to the property of the elastic foundation at the bottom of the models, we performed an additional series of experiments with the 40-km-long fault, in which we varied the density of the lower crust between 2800 and 3100 kg/m³. The results show that the slip rate increase induced by surface processes does not depend on the property of the elastic foundation for a reasonable density of the lower crust, i.e. the slip curves in the additional and the original models are almost identical (not shown in figure).

To illustrate how the elevation of fault footwall and hanging

wall have evolved after 750 ka in the experiments with and without erosion and deposition, we extracted topographic profiles from the experiment without surface processes and with variable diffusion constants, respectively (Figs. 2, 8a). The profiles run across the centre of the 40-km-long, 60°-dipping normal fault at the model surface and at 2 km depth, i.e. along the top surface of the tetrahedral element layer. Without surface processes, footwall uplift creates a ridge with the highest point at an elevation of 218 m located close to the fault, whereas the deepest point of the hanging wall has subsided to a depth of 228 m (Fig. 8a). If erosion and deposition are active in the model, the elevation of the footwall decreases with increasing diffusion constant while the highest point migrates away from the fault. This shift of the maximum surface uplift away from the fault is also revealed by comparing the spatial pattern of the vertical displacement rate derived from the experiments with and without surface processes (Fig. 8b). The cross-sections through the centre of the fault further show that the subsidence rate of the hanging wall increases if it is affected by the deposition of sediments (Fig. 8b).

Fig. 9. Rates of erosion and sedimentation and the induced stress field in the crust derived from the experiments with a 40-km-long, 60°-dipping normal fault (cf. Figs. 2, 3). The two upper diagrams show the temporal evolution of a) erosion of the footwall and b) sedimentation on the hanging wall of the fault as derived from the experiments with variable diffusion coefficients k_D (Fig. 2). The amount of erosion and sedimentation was averaged spatially over a 20 km×40 km large area in the fault footwall and hanging wall, respectively (see inset maps). c) Schematic sketch of the stress field in the model crust around the normal fault as derived from an experiment with surface processes. Owing to sediment deposition in the fault hanging wall, the vertical stress S_v increases near the fault. This leads to an increase in the maximum principal stress (σ_1), which is rotated toward the fault plane in the vicinity of the fault, and hence to a local increase in the differential stress ($\sigma_1 - \sigma_3$). The increase in the differential stress causes the slip acceleration on the fault. Note that the rotation of the maximum principal stress σ_1 is predominantly caused by the slip on the fault; in the presence of surface processes, the cumulative rotation is $\sim 3^\circ$ higher at the end of the model time.

The amount of erosion averaged over a 20×40 km large region in the fault footwall is 14, 19 and 24 m at the end of the experiments with diffusion constants of 0.1, 0.3 and 3 m²/a, respectively (Fig. 9a). The rate of erosion gradually increases with time and reaches 0.025, 0.032 and 0.042 mm/a for the different diffusion constants. These modelled erosion rates are similar to those determined by cosmogenic nuclides in low-relief mountain ranges (e.g. Meyer et al., in press). The sediment thickness averaged over an equally

large area on the hanging wall amounts to 9, 15 and 22 m at the end of the experiment (Fig. 9b), with rates ranging from 0.016 to 0.038 mm/a for increasing diffusion constants. Subsidence and flexure of the hanging wall increase in the presence of surface processes owing to the weight of the sediment deposited on the hanging wall, which is most obvious when comparing the profiles at ~ 2 km depth. The profiles across the model surface show that the thickness of the sediment deposited in the hanging-wall basin is

Fig. 10. Results of a model including three normal faults of different length after 750 ka of extension at a rate of 2 mm/a. **a)** Perspective view of the horst and basins generated by the slip on the three faults. The diffusion constant in this experiment is $k_D = 0.1 \text{ m}^2/\text{a}$ and the fluvial erosion parameter $K_f \cdot p = 0.03 \text{ m/a}$. Vertical exaggeration is 5. **b)** Temporal evolution of slip on faults 1–3.

considerably higher than the amount of flexure and subsidence of the hanging wall only in the experiment with a diffusion constant of $3 \text{ m}^2/\text{a}$ (Fig. 8a). It is this mass redistribution in the fault footwall and hanging wall that ultimately leads to the slip rate increase on the fault by affecting the stress field of the crust. The experiment without surface processes shows that the differential stress ($\sigma_1 - \sigma_3$), which ultimately controls the slip behaviour of the fault, increases with time in the crust because of the ongoing horizontal extension. If erosion and sedimentation are active, this increase in the differential stress is more pronounced near the fault and therefore induces a higher fault slip rate. The ultimate reason for the change in the differential stress is an increase in the vertically oriented normal stress component (S_v) near the fault, which results from the deposition of sediments in the hanging wall (Fig. 9c). This increase in the vertical normal stress component (S_v) leads to an increase in the differential stress because it mainly affects the maximum principal stress σ_1 , which is rotated toward the fault plane in the vicinity of the fault. Note that the rotation of the maximum principal stress σ_1 is predominantly caused by the slip on the fault; in the presence of surface processes, the rotation is increased. By the end of the model time (750 ka), the maximum principal stress (σ_1) has rotated by $20\text{--}25^\circ$ when surface processes are absent whereas in the presence of surface processes this cumulative rotation is $\sim 3^\circ$ higher. In all experiments with surface processes the increase in differential stress relative to the experiments without surface processes is similar to the slip rate increase, e.g. $\sim 15\%$

in the model with a 40-km-long, 60° dipping fault.

Our experiments do not only reveal how fault slip at depth is influenced by the surface processes on the model surface but also how erosion and deposition are affected by the active fault. Driven by the horizontal extension, the model fault accommodates slip with an arc-shaped distribution along strike, which mimics displacement profiles of natural normal faults (Cowie and Scholz, 1992a,b,c; Schlische et al., 1996; Cowie and Roberts, 2001; Manighetti et al., 2001; Roberts and Michetti, 2004). In other words, the amount of accumulated slip is largest at the fault centre and diminishes toward the fault tips. As a result, the maximum footwall uplift and hanging-wall subsidence also occur at the fault centre. The topography created by the fault determines the pattern of erosion and sedimentation acting on the model surface. Some streams draining the footwall flow around the fault tips or across the escarpment and provide sediment to the hanging-wall basin (Fig. 2). Other streams incise the uplifting footwall and due to its shape some of them flow away from the fault, carrying sediments outside of the model. The rate of total material outflow across the model boundaries at the end of the experiments is governed predominantly by the product of transport coefficient and precipitation rate. It increases from $2300 \text{ m}^3/\text{a}$ for $K_f \cdot p = 0.003 \text{ m/a}$ to $89,800 \text{ m}^3/\text{a}$ for $K_f \cdot p = 0.3 \text{ m/a}$. In contrast, the material outflow is similar ($19,400 \text{ m}^3/\text{a}$) in the three experiments with variable diffusion constants.

4.2. Applicability of model approach to more complex tectonic settings

To demonstrate the wide applicability of our model approach and to show that the findings described above remain valid also for more complex tectonic settings, we present an additional model with a horst generated by slip on three 60°-dipping normal faults of different lengths (Fig. 10). In such a fault array, the slip evolution of each fault is primarily controlled by the horizontal extension rate (2 mm/a), the fault geometry and the interaction with the neighbouring faults (e.g. *Schlische and Anders, 1996; Schlische et al., 1996; Gupta and Scholz, 2000*). As revealed by the experiments with a single fault, however, the interaction of tectonic and surface processes may also affect the slip behaviour of the faults.

The results of this experiment show that the spatial distribution of the fault-generated uplift and subsidence controls the stream direction and impose a more complex erosional and depositional pattern than in the experiments with an isolated normal fault. The rising horst is drained by streams that flow toward the subsiding hanging walls or leave the model at the boundaries. Note that one of these streams drains part of the uplifted area and flows through the ramp between the en-echelon arranged normal faults. This stream delivers additional sediments to the hanging wall of fault 1, which form a delta in the hanging-wall basin (Fig. 10a). As a consequence of this additional deposition, the slip rate of fault 1 increases by 11.2% compared to the control run without surface processes, which is a higher slip rate increase than that obtained from the isolated normal fault of the same length (8.8%; Table 1). In contrast, the hanging wall of fault 2, which is also 20 km long, does not experience such additional, irregular sedimentation and hence its slip rate increases only by 8.6% compared to the slip rate without surface processes, i.e. the slip rate increase is similar to that of an isolated 20-km-long fault (Table 1). The slip rate on the 40-km-long fault 3 increases by 9.8% in the presence of surface processes, an increase close to that of an isolated normal fault of the same length (Table 1). In summary, the horst experiment shows that the intensity of the feedback between surface processes and tectonics may vary spatially depending on the distribution of areas experiencing erosion or deposition and their position in the fault array.

5. Conclusions

Using a novel approach that couples 3D tectonic models including discrete faults with a landscape evolution model, we showed that erosion and sediment deposition affect the slip behaviour of isolated normal faults by accelerating their slip rate by up to ~15%. The magnitude of the slip rate increase caused by surface processes depends predominantly on the diffusion constant as well as on the length and dip of the fault. An experiment with an array of three normal faults creating a horst shows that the slip rate increase caused by erosion and deposition also occurs in more complex tectonic settings. Our results demonstrate that a positive feedback exists between tectonics and surface processes, which leads to an additional amount of fault slip caused by erosion and sedimentation. More generally, our study demonstrates the great potential of the new modelling tool and its applicability to complex tectonic settings, which will open new horizons for deciphering the interaction between faulting and surface processes.

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