Interplay between lithospheric flexure and river transport in foreland basins

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ABSTRACT
Foreland basins form by lithospheric flexure under orogenic loading and are filled by surface transport of sediment. This work readdresses the interplay between these processes by integrating in a 3D numerical model: the mechanisms of thrust stacking, elastic flexural subsidence and sediment transport along the drainage network. The experiments show that both crustal tectonic deformation and vertical movements related to lithospheric flexure control and organise the basin-scale drainage pattern, competing with the nonlinear, unpredictable intrinsic nature of river network evolution. Drainage pattern characteristics are predicted that match those observed in many foreland basins, such as the axial drainage, the distal location of the main river within the basin, and the formation of large, long-lasting lacustrine systems. In areas where the river network is not well developed before the formation of the basin, these lithospheric flexural effects on drainage patterns may be enhanced by the role of the forebulge uplift as drainage divide. Inversely, fluvial transport modifies the flexural vertical movements differently than simpler transport models (e.g. diffusion): Rivers can drive erosion products far from a filled basin, amplifying the erosional rebound of both orogen and basin. The evolution of the sediment budget between orogen and basin is strongly dependent on this coupling between flexure and fluvial transport: Maximum sediment accumulations on the foreland are predicted for a narrow range of lithospheric elastic thickness between 15 and 40 km, coinciding with the $T_e$ values most commonly reported for foreland basins.

INTRODUCTION
Foreland basins are depressions formed by bending of the lithosphere under the weight of a migrating orogenic wedge. Their sedimentary infill is the result of surface mass transport from the orogen to the subsided regions (Fig. 1), and records the interaction between deep (lithospheric) and surface processes. This paper focuses on the interplay between the flexural behaviour of the lithosphere and the sediment transport driven by the fluvial network.

Lithospheric flexure and mass balance in foreland basins

The concept that lithosphere behaves as a flexible thin plate when it is submitted to external forces has been tested in many different geological settings such as seamounts (Vening-Meinesz, 1941) and oceanic trenches (e.g. Bodine et al., 1981; Garcia-Castellanos et al., 2000). Since the early quantitative studies of foreland basin formation (Beaumont, 1981; Jordan, 1981), it is generally accepted that the sedimentary record of these basins is related to the flexural response of the continental lithosphere to orogen loading. Later numerical models incorporated non-instantaneous wedge loading and sedimentation processes (Flemings & Jordan, 1989; Toth et al., 1996; Ford et al., 1999) finding that larger sedimentary basins are expected at higher transport rates and higher lithospheric rigidity (Flemings & Jordan, 1989). Other authors investigated the role of the lithospheric rheological stratification on the geometry of the sedimentary infill (Waschbusch & Royden, 1992; Garcia-Castellanos et al., 1997, 2002). Although basin stratigraphy is dependent on the rheological model adopted for the lithosphere (elastic, viscous, plastic, or a combination of them), the geometry of the basement can be generally modelled as an elastic thin plate loaded by thrusting and occasionally by deeper processes such as subduction and slab detachment (Royden, 1993; Sinclair, 1997). Besides these links between flexure and surface transport, quantitative models have shown that erosion exerts also a strong control on crustal (Beaumont et al., 1994, 2000; Willet, 1999) and lithospheric (Avouac & Burov, 1996) orogenic deformation (Fig. 2).

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Fig. 1. Sketch showing the elements incorporated in the conceptual model of foreland basin formation: Thrust stacking produces bending and subsidence of the lithosphere and forebulge uplift of the foreland. Surface mass redistribution is driven by river transport.

Fig. 2. Interplay between the main processes involved in foredeep basin evolution. Dotted arrows indicate processes not explicitly accounted for in this work.

(c.g. constant rate of erosion/deposition, diffusion, etc.), failing to account for the transport dynamics of the river network and its remarkable three-dimensionality. In fact, fluvial transport plays a main role in determining the asymmetries of the basin geometry and the depositional environments (e.g. Johnson & Beaumont, 1995; Schlunegger et al., 1997). Furthermore, some studies note that most basin subsidence is related to the weight of the sediment rather than the tectonic generation of accommodation space (e.g. Schlunegger et al., 1997). This opens the question mark of whether more realistic transport models can predict large deposits in front of the orogen in absence of relevant flexural subsidence, and points to the necessity of a systematic study of the role of fluvial transport in foreland basin formation models.

Drainage and transport in foreland basins

Because foreland sedimentary basins are relief depressions next to orogens, they usually collect important amounts of rainwater, forming first order hydrological basins. Although drainage patterns of foreland basins are very diverse, a large set of them share two conspicuous features: Drainage is axial (parallel to the orogen) and the main river runs along the passive (external or distal) margin of the basin. Geological examples are sketched in Fig. 3. The Danube River runs along the northern margin of the North Alpine Basin, very close to the Hercynian basement outcrop. About 1000 km downstream, the same river repeats this pattern in the South Carpathian Basin. The Guadalquivir River (Spain) drains the Guadalquivir Basin and the Betic Range along the northern external boundary of the basin. The Ganges River and its main tributaries, after draining the Himalayas, cross perpendicularly the entire foreland basin and then run eastwards about 1000 km along the contact between the basin and the Indian basement outcrop (Burbank, 1992). Another example is the Paraguay River, which flows close to the Palaeozoic foreland outcrop, along the featheredge of the Chaco Basin.

This external location of the main river draining foreland basins is peculiar since the concept of foreland basin formation suggests a topographic minimum (foredeep) next to the orogenic wedge, in the active rather than the passive margin of the basin. The distal location of the Ganges River within the Ganges foreland basin has been related to the tectonic evolution of the Himalayas and to the sediment supply from this orogen (Burbank, 1992): lower tectonic shortening rates and/or higher orogen erosion would facilitate, within this model, a distal location of the main river by uplifting and tilting the basin and increasing the transversal sediment delivery (Heller et al., 1988).

Since rivers are typically responsible for most of the mass transport from orogen to basin, their dynamics are important to understand the evolution of foreland basins (e.g. Gupta, 1997). Although the physics of river transport are complex and still not fully understood, simple approaches permit to explain general characteristics of river profiles and patterns of drainage basins. Different models have been proposed relating the river transport capacity to water discharge, riverbed slope, sediment grain size, and rock erodability (Willgoose et al., 1991; Beaumont et al., 1992; Seidl et al., 1992; Howard, 1994; Kooi & Beaumont, 1994; Whipple & Tucker, 1999). These models have been used extensively to improve our understanding of landscape and drainage evolution and the large-scale geomorphological implications of river erosion (Kooi & Beaumont, 1994, 1996; Braun & Sambridge, 1997; Tucker & Slingerland, 1997; Kühni & Pfiffner, 2001), proving that simple formulations account for various geomorphological concepts such as base level, steady state equilibrium landforms, peneplanation, escarpment retreat, river dynamic equilibrium, river time response, etc. Tucker & Slingerland (1996) used these techniques to address the impact of tectonism on orogen drainage patterns and sediment supply from fold-and-thrust belts. Johnson & Beaumont (1995) show preliminary results of a 3D numerical model incorporating flexure and fluvial transport, and suggest that the longitudinal drainage may be conditional to laterally diachronous development of the orogen. However, their model does not account for the role of the initial topography or the formation of lakes in local topographic minima.
The main rivers draining each basin are indicated. Except for the Po Basin, which is tectonically loaded from two directions, all these examples present a distal location of the main river.

**Aim**

From the considerations above, fundamental questions demand a systematic analysis of a 3D quantitative model of foreland basin formation incorporating fluvial transport: How does flexure affect the drainage pattern in a foreland basin? What is the role of river transport on the history of sediment accumulation? In this sense, the purpose of this work is to address the interplay between lithospheric flexure and river networks during the 3D mass transport at foreland basins by developing a numerical algorithm that integrates quantitative approaches to these processes.

**QUANTITATIVE MODELLING OF FORELAND BASIN FORMATION**

The mass balance between orogen erosion and basin sedimentation is a key constrain for foreland basin modelling. Since this balance is driven typically by longitudinal sediment transport, a two-dimensional (cross section) modelling approach is inadequate. The conceptual model designed in this work considers the 3D interactions between three processes (Fig. 1): (1) Crustal scale thrusting related to crustal shortening; (2) Flexural response of the lithosphere to mass redistribution; and (3) Large scale river erosion, transport and sedimentation of sediments. Although deeper processes (e.g. plate subduction) have been invoked to explain the excess or deficit of subsidence predicted from the topographic load in many foreland basins (e.g. Royden & Karner, 1984; Bott, 1991) their effect has been disregarded in this study for simplicity, assuming that the load contributing to basin subsidence is related only to (upper) crustal mass redistribution.

**Fluvial network**

The transport properties of rivers depend strongly on water discharge and riverbed slope (e.g. Strahler, 1978). Since river discharge increases downstream due to the confluence of tributaries (Hack, 1957), it is very sensitive not only to the rainfall distribution but also to the particular organization of the topography upstream. For this reason, it is necessary to incorporate to the model the process of river network organization, and directly relate the amount of erosion to independent observables such as rainfall (runoff) distribution.
Each location in the model has a certain water discharge as a result of the addition of local runoff (rainwater delivered to the drainage network) and water coming from the tributary surrounding nodes. The resulting discharge is transferred to the lower node of the river (determined by the maximum slope) after evapotranspiration has been subtracted. The water balance in each node can be written as (Fig. 4)

\[
\text{output water} = \sum \text{tributary inputs} + \text{local runoff} - \text{evapotranspiration.}
\]

Conservation of total water flowing all over the model at a given time must also be accomplished:

\[
\text{total rain} - \text{evapotranspiration} = \text{water drained to the sea} + \text{water flow at boundary.}
\]

Formation of lakes in local topographic minima is accounted for in the model because the geometry of lakes plays a major role in drainage evolution, since it determines the localization of sedimentation in the water body and erosion and flush in the outlet.

**Sediment transport**

River dynamics studies have found that river incision has a complex dependence on a large set of parameters including water discharge, riverbed slope, sediment grain size, and rock erodability (e.g. Howard, 1994; Tucker & Slingerland, 1997; Whipple & Tucker, 1999). Because the purpose here is not the study of the processes involved in fluvial incision, the open discussion on the best analytical relationships representing those processes (e.g. van der Beek & Braun, 1999; Whipple & Tucker, 1999) is not addressed in this paper. The approaches of previous models of drainage basin evolution by Beaumont et al. (1992) and Kooi & Beaumont (1994) are adopted. Following these authors, river transport can be approached by means of the concept of sediment transport capacity in equilibrium, defined as the amount of mass transported by a river wherever it is in equilibrium (i.e. produces no net erosion or sedimentation). This capacity is observed to be proportional to the mean water discharge \( Q_w \) (measured in units of volume/time) and the slope \( S \) of the river profile, and the simplest linear relationship is adopted by some authors (e.g. Kooi & Beaumont, 1994):

\[
q^d(x, y, t) = K_f \cdot S(x, y, t) \cdot Q_w(x, y, t)
\]

where \( K_f \) is the fluvial transport coefficient, which can be obtained by matching measurements of mean sediment load and water discharge. The value used in all the models shown in this paper (60 kg/m³) is within the range of values used by other authors (e.g. Kooi & Beaumont, 1996; van der Beek & Braun, 1999). For comparison with these works, note that here the sediment load \( q \) is measured in units of mass/time, rather than volume/time, which modifies the units of \( K_f \). In general, rivers are out of equilibrium and then the amount of material \( dq \) eroded or deposited along a river segment of length \( dl \) is approached as proportional to the difference between the actual sediment load \( q \) and \( q^d \) following the equation (Beaumont et al., 1992)

\[
\frac{dq(x, y, t)}{dl} = - \frac{1}{l_e} \left( q(x, y, t) - q^d(x, y, t) \right)
\]

where \( l_e \) is the length scale of erosion/deposition. Within this approach, a river changes from incision to aggradation by a reduction in its capacity \( q^d \), i.e. by a decrease in discharge and/or slope.

Mass conservation (sediment budget) in each cell of the model (see Fig. 4) can be written as

\[
\text{output sediments} = \sum \text{tributary sed. Inputs} + \text{local erosion} - \text{local sedimentation.}
\]

The total amount of erosion in the model domain is either deposited or gone through the boundaries of the model:

\[
\text{eroded mass} = \text{deposited mass} + \text{mass outflow at boundaries.}
\]

Sedimentation of the material transported to lakes and sea is simulated by distributing the sediment in all directions from the river mouth and deposited assuming null transport capacity in Eq. (2). This implies a deposition rate decreasing exponentially with the distance from the
Loadural formation

Flexural isostatic vertical motions are related to mass redistribution all over the model. The flexural load increment at each time step is defined as the increment in weight at every column of the model and is related to the mechanisms of crustal shortening, erosion, sediment accumulation, and water layer thickness.

Because this work is not intended to study neither the internal configuration of the orogenic wedge nor the dynamics of orogen growth, orogenic growth is considered as a kinematic process generating non-instantaneous thrust load and relief. The geometry and velocity of the thrusting blocks must be therefore predefined for each model. Movement of each block along thrusts is quantified by means of the vertical shear approach, i.e. shortening is accommodated along one or several thrust faults by moving the upper block while preserving its vertical thickness (e.g. Toth et al., 1996). This approach allows calculating the load increment at each location as a function of the amount of displacement and the block geometry.

Sediments are deposited on the top of the model and dragged by the thrusting basement units below. Because the internal geometry of the tectonically deformed sediments is out of the scope of this study (its typical scale is one or two orders of magnitude smaller than the basin), sediment thickness is duplicated in the nodes in front of the orogenic wedge without keeping track of the internal fault geometry. This approach concentrates deformation in front of the active thrust and accomplishes the conservation of mass. The weight of the eroded (or deposited) material at each node is subtracted (or added) to the total load resting on the flexural plate.

Flexural subsidence

Vertical isostatic movements of the lithosphere are of primary importance during drainage development, since they partly regenerate the eroded topography by erosional rebound (e.g. Tucker & Slingerland, 1994) and provide the necessary space for sediment accumulation. To account for the isostatic compensation of the mass redistribution occurring during the basin evolution, a 2D (planform) thin elastic plate approach is used. A fourth order differential equation (e.g. van Wees & Cloetingh, 1994) relates the flexural elastic deflection related to a 2D load distribution. Young modulus adopted value is $E = 7 \times 10^{10} \text{N/m}^2$ and Poisson's ratio is $\nu = 0.25$. The flexural equation governing the behaviour of the plate is linear in respect to load, which permits to calculate successive increments of deflection as resulting from increments of load.

Numerical model

The mathematical approaches described above were integrated in a single computer program written in C language. In this forward model, for each time step a sequence is repeated in which tectonic deformation, fluvial transport and lithospheric flexure are calculated. All the equations discussed in the previous section have been solved using the finite difference technique in both space and time. In the case of the tectonic deformation, the use of a rectangular mesh of about $5 \times 5 \text{km}$ cells imposes the limitation that moving blocks must advance an integer number of nodes at every time step. For this reason, both flexure and tectonic deformation are solved with time steps between 0.5 and 0.78 Myr, one order of magnitude larger than those used for surface transport. To avoid abrupt changes of the topography, the calculated effect of flexure and tectonism is applied gradually during the surface transport iterations. Tests undertaken with smaller time steps and smaller spatial discretization did not introduce qualitative changes in the results.

In order to incorporate the process of lake sediment infill, the algorithm identifies the extension of lakes and their outlets on the dynamic topography. For simplicity, evapotranspiration is assumed null in the models shown in this work, which implies that lakes are open (have at least one outlet) and the discharge delivered through the outlet is equal to the addition of water inputs from the tributary rivers plus the total rainfall in the lake. Assuming null evapotranspiration implies also that all the precipitation is collected as runoff.

The boundary conditions adopted for the flexural subsidence consist on imposing (1) null derivative in the direction perpendicular to the boundary, and (2) null curvature across the boundary. In physical terms, this is equivalent to assume that no external torque is applied to the plate. For the surface processes, I assume that a node in the boundary having no lower surrounding nodes drains out of the model with a slope equivalent to its major tributary. This means that the erosion/sedimentation rate is approximately equal to that in the inner vicinity. Finally, tectonic movement of an incoming body assumes that the body thickness is the same beyond the boundary, and therefore its thickness at the boundary is not modified by its displacement.

Example: Regular spacing of drainage outlets

As an example of the surface fluvial transport in absence of tectonism, we use an initial linear escarpment dividing a 1000-m high peneplane to the right from a 2000-m deep ocean to the left (the initial shore line is at $x = -5 \text{ km}$). Other parameters are listed in Table 1. The plane in the top behaves initially as an area of unorganised drainage.
basement density small dominated by shallow lakes, which form because of the drainage at initial topography (dashed line) and the final prograding geometry of the sediments. Parameters are listed in Table 1.

| Initial time | 0 Myr |
| Final time   | 15 Myr |
| Time interval dt transport | 0.05 Myr |
| Sediment density | 2200 kg m⁻³ |
| Basement density | 2800 kg m⁻³ |
| Precipitation | 300 mm y⁻¹ |
| Fluvial transport coef. $K_f$ | 60 km⁻¹ |
| Length erosion $l_f$ | 120 km |
| Length erosion of sediments $l_{fr}$ | 60 km |
| Length $l_s$ sedimentation | 25 km |
| Gridding $dx,dy$ | 10 x 10 km |
| Initial topog. random noise | 10 m |
| Initial topography | -2000 to 1000 m |

Hovius (1996) and Talling et al. (1997) found a good correlation between outlet spacing and length of fluvial basins draining linear geological structures. The ratio between these two observables is restricted to a narrow range between 1.91 and 2.23 for a set of 11 mountain belts and between 1.4 and 4.1 in fault blocks. This remarkable regularity, for which up to date there is no proposed physical basis, is thus valid in a wide range of scales. In the model displayed in Fig. 5, the number of outlets corresponding to basins that reach the eastern limit of the model is 5, with a mean spacing of $S = 30$ km. Since the width of the drainage basins is $W = 75$ km, the ratio $R = W/S$ is 2.25, a realistic value in agreement with the observations by Hovius (1996) and Talling et al. (1997). The effect on $R$ produced by variations of model parameters such as gridding spacing (model resolution), length scale, precipitation and transport coefficient is relatively small, obtaining a range of $R$-values between 1.4 and 2.9. The main factor controlling the shape of the drainage basins is the initial topography and vertical movements, since they determine the initial drainage. For example, by incorporating to this model either the flexural isostatic rebound or a moderate initial slope towards the right, the initial escarpment becomes a water divide and the timing of escarpment retreat is slowed down by a factor 2 or higher. This result coincides with those obtained by Tucker & Slingerland (1994).

**FLEXURE-DRAINAGE INTERPLAY DURING SURFACE MASS REDISTRIBUTION – FRONTAL CONVERGENCE MODEL**

The interplay between regional isostasy and fluvial surface transport is first studied by means of a synthetic setting in which crustal shortening occurs perpendicular to an initial continental margin (frontal convergence model). Figure 6 shows four stages of the evolution obtained for this model. The initial topography ($t = 0$ Myr) corresponds to a central sea up to 500 m deep limited by two 300 m high peneplanes to the North and South. Convergence occurs along two linear, inverse faults with a 10-km deep detachment level (Fig. 6a). These faults strike E–W and reach the surface at $y = 20$ km and $y = -35$ km, respectively (Fig. 6b). Southwards directed thrusting occurs with a velocity of 5 km/Myr between 0 and 6 Myr (along the northern fault) and between 6 and 12 Myr (along the southern fault). Other parameters are given in Table 2.

Four main phases can be distinguished in the drainage evolution of the frontal convergence model (Fig. 6b): (1) ($t = 0$ Myr) Initial unorganised drainage in the peneplane; (2) (0–12 Myr) Marine basin formation due to foreland flexural subsidence: Division of drainage by the forebulge; (3) (12–17 Myr) Transitory post-tectonic lake depositional systems; and (4) (after 17 Myr) Continental fluvial
Fig. 6. Frontal convergence model: (a) Schematic N-S cross-section of the initial setting indicating the geometry of the two successive thrusts; (b) Evolution of the model. Each panel shows the topography illuminated from the north (contours at -700, -200, 0, 200, 600, 1000, 2000 and 3000 m above sea level) and the drainage (river thickness proportional to water discharge). Model parameters are listed in Table 2. The initial stage \( t = 0 \) shows the front of the two moving blocks (red dashed lines). Subsequent stages are representative of the underfilled period \( t = 6 \) Myr, the lacustrine period \( t = 14 \) Myr and the filled period \( t = 24 \) Myr.

Table 2. Parameters used for the frontal convergence model. Other parameters as in Table 1.

| Frontal convergence model | \( T_e \) | km | Initial time | 0 | Myr | Final time | 24 | Myr | Time interval for tectonics | 0.5 | Myr | Time interval for transport | 0.05 | Myr | Shortening velocity | 5 | km·Myr\(^{-1} \) | till \( t = 12 \) Myr (N-S) | 2200 | kg·m\(^{-3} \) | Sediment density | 2800 | kg·m\(^{-3} \) | Precipitation | 300+ | mm·y\(^{-1} \) | Gridding \( N_x, N_y \) | 101 × 101 | km | Width cell \( dx, dy \) | 5 × 5 | km | Initial topography | -500 to 200 | m

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deposition in the foreland basin. Drainage becomes axial and the main river shifts towards the foreland.

The last stage \( t = 24 \) Myr; Fig. 7) shows the post-tectonic well-developed drainage network, eroding more efficiently in the orogen than in the basin due to the higher slope. The main rivers are located next to the passive (external or distal) margin of the two sedimentary basins, collecting the runoff of the precipitation from the orogen, the basins and the inner side of the forebulge uplift.

Axial drainage is clearly the dominant situation in most tested models. Since the early development of the axial drainage, the main river is located distal because of the higher sediment supply from the transversal tributaries of the hinterland relative to those coming from the foreland. This higher sediment supply is due to the higher relief of the orogen (increasing the slope in Eq. (1)) and its orographic effect on precipitation (increasing water discharge in Eq. (1)).

The model tracks the distribution of erosion and sedimentation and the sediment balance through time (Fig. 8a). It is of particular interest how the total volume of sediments in the model depends on \( T_e \), keeping all other parameters fixed. Figures 8(a) and 8(b) show the
between 15 and 40 km. As seen from earlier foreland basin basins require longer time to fill after the cessation of boundary, i.e. towards the E and W limits of the initial sea. These areas do not suffer tectonic deformation but deeps) at the end of the tectonic deformation, and these subsidence because of their proximity to the orogen. The accommodation space that lead to very underfilled basins (fore- during forebulge formation is now further studied by the Betics-Guadalquivir basin (oblique convergence model). In order to keep this synthetic setting grounded on a geological example, this model has many (intentioned) similarities with the Betics-Guadalquivir system (Fernández et al., 1998; García-Castellanos et al., 2002) located in Fig. 3. The imposed lateral variation of $T_c$ ranges from 5 km (east boundary) to 20 km (west precipitation rate relative to the previous model. Although erosion increases in about 50–100%, the basin sediment volume is smaller, particularly during the post-tectonic phase, because the faster erosional removal of topographic load reduces flexural subsidence and increases post-tectonic flexural rebound and basin incision (Heller et al., 1988). Only at this high erosion rates, and for larger orogen lengths, can the basin eventually become overfilled and the transversal rivers overcome the forebulge barrier and drain towards the foreland.

To show that these features are intrinsic to fluvial transport, diffusive surface transport was applied to the frontal convergence model designed above. The results show (Fig. 8d) that the diffusive transport model predicts large basin volumes for high $T_c$ values, in agreement with the results by Flemings & Jordan (1989), but in contrast with what is predicted above assuming fluvial transport. Since the tectonic setting has nearly a 2D symmetry, the diffusive model predicts the deposition of all eroded material next to the orogen, similar to Flemings & Jordan (1989), whereas in the fluvial transport model the axial drainage delivers most of the erosion products out of the model domain after the basin is filled.

**ROLE OF FLEXURE ON FORELAND BASIN DRAINAGE – OBLIQUE CONVERGENCE MODEL**

The interplay between flexure and drainage networks during forebulge formation is now further studied by means of a second end member synthetic case in which convergence occurs parallel to the initial margin (oblique convergence model). In order to keep this synthetic setting grounded on a geological example, this model has many (intentioned) similarities with the Betics-Guadalquivir system (Fernández et al., 1998; García-Castellanos et al., 2002) located in Fig. 3. The imposed lateral variation of $T_c$ ranges from 5 km (east boundary) to 20 km (west
boundary), reflecting the values obtained by van der Beek 
& Cloetingh, 1992) and Garcia-Castellanos et al. (2002) for 
the Guadalquivir Basin (South Spain). The tectonic kinematics, though very simplified from a structural point of 
view, reflect the incoming of the Alborán Domain (emplaced from east to west) and the crustal thickening below the Betics. A constant E–W shortening rate of 7 km/ 
Myr is accommodated by four arched thrusts with a dip angle of 10° and an 11-km deep detachment level. These 
faults activate successively from E to W between \( t = 0 \) and 
\( t = 22 \) Myr. Other parameters are given in Table 3.

Table 3. Parameters used for the oblique convergence model.

Other parameters as in Table 2.

| Lithospheric flexure and fluvial transport
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<td><strong>Oblique convergence model</strong></td>
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<td>( T_e )</td>
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<td>Final time</td>
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<td>Time interval for tectonics</td>
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<td>Time interval for transport</td>
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<td>Shortening velocity</td>
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<td>Initial topography</td>
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Figure 9 shows four stages of the drainage evolution obtained for the oblique convergence model. The initial 
topography \( (t = 0 \) Myr) corresponds to an E–W striking continental margin, which is \( 300 \) m high on-shore and 
\( 1700 \) m deep offshore, varying linearly between both 
values along a 110-km wide transition stripe.

In contrast with the frontal convergence model, here 
only one subaerial foreland basin is formed because the 
initial conditions imply a submarine topography in the 
retro-side of the orogen. Because the initial configuration 
in this model facilitates the drainage towards the sea (due to 
the obliquity between the initial geometry and the load 
stacking), only three main phases are distinguished in its 
drainage evolution: (1) \( t = 0 \) Myr Initial unorganised 
drainage in the continent peneplane; (2) \( 0–22 \) Ma Mostly 
marine deposition related to sea invasion of the subsided 
foreland. The uplifted forebulge becomes a drainage divide 
between the foreland and the basin. The eastern side of the 
basin becomes subaerial as deposition progradates towards 
the W. Drainage becomes axial in the subaerial area and 
the main river is ‘expulsed’ towards the foreland; and (3) (after 
\( 22 \) Ma) Post-tectonic, mainly continental deposition in the 
filling foreland basin, drainage remains axial and distal. 
Note that in this model, in contrast with the previous 
frontal convergence model, the distal, axial drainage 
forms already during the syn-tectonic stage.

The final configuration of the system (Fig. 10) shows an important sediment accumulation on the foreland related to the flexural subsidence (the sediment thickness on the initially subaerial basement reaches 4000 m). As in the case of frontal convergence, a remarkable characteristic is the axial drainage along the basin axis and the location of the main river next to the passive (northern) margin of the sedimentary basin. This river collects the water discharge from the sedimentary basin itself, but also from the northwestern slope of the mountain range and the southern flank of the forebulge. This final configuration of drainage and basin geometry reproduces many features of the Guadalquivir Basin (Fig. 3). Apart from the correct prediction of the location of the main river (Guadalquivir River) in the basin, the model drainage divide induced by forebulge uplift coincides with the drainage divide in Sierra Morena, supporting the idea that the relief of this mountain range is mainly related to forebulge uplift (Garcia-Castellanos et al., 2002).

Erosion rate increases during the periods of tectonic activity (Fig. 11), showing a delay between tectonism and fluvial erosional response (Kooi & Beaumont, 1996). As tectonic deformation stops, the erosion rate decreases exponentially to zero. The lithospheric elastic thickness \( T_e \), together with the transport parameters, controls the rapidity of this decrease, because it determines the isostatic restoration of the eroded topography: low \( T_e \) values are more efficient for this erosional isostatic rebound, implying slower decays of the erosion rate. In connection to this uplift, the sedimentary infill of the basin becomes progressively affected by erosion during the post-tectonic stage, first in the proximal parts of the basin. Eventually, the amount of erosion exceeds the deposition and the basin begins decreasing in volume (\( T_e = 29 \text{ Myr} \)).

To check whether flexure is relevant in the resulting drainage pattern, the evolution of the oblique convergence model has been recalculated assuming a local isostasy (\( T_e = 0 \); results in Fig. 12a) and absence of isostasy (no vertical movements or \( T_e = \infty \); Fig. 12b). These results, coinciding with those obtained from the frontal convergence model, show that neither \( T_e = 0 \) nor \( T_e = \infty \) allow to predict basin formation on initially subaerial areas, since the only important sediment accumulations occur in areas that were initially deep below sea level. In addition, the basin drainage patterns obtained in the models above are now vanished because of the lack of vertical movements in the foreland.
EFFECT OF SMALL INITIAL PERTURBATIONS

The coupled tectonic-surface transport model is chaotic, in the sense that small perturbations of the initial conditions lead to large changes in the results. Many modified versions of the settings described above were tested to ensure that the achieved conclusions have general validity in spite of the non-linear model dynamics. For example, the effect of 1 m increase of the depth of the first detachment fault (only 0.01% of the original depth) is dramatic in the sense that the particular location of each stream is modified at $t = 30$ Myr. A similar effect is introduced by the numerical errors during the computation, explaining why the initial perfect E–W symmetry of the frontal convergence model is not maintained during the model evolution (Fig. 6).

This is better illustrated by adding a random noise to the flat initial topography of the foreland ranging from -30–30 m. The results in Fig. 13 show that the particular location of the rivers is modified relative to the unperturbed model, but not the general characteristics of the model such as the volume of sediments retained in the basin or the main basin drainage patterns (i.e. the axial drainage, the forebulge drainage divide and the distal location of the main river).

Small perturbations predict relevant differences in topography in the orogen but only of very small wavelength, whereas the regional topography (large wavelength) is maintained (Fig. 13a). Increased random perturbations of the initial topography overprint the role of the forebulge as the drainage divide in the foreland. Perturbations higher than c. 50 m virtually obliterate the forebulge effect on the fluvial network because the initial drainage organization becomes more important than the subsequent vertical motions. This suggests that the effect of flexural forebulge uplift on drainage is in general irrelevant except for settings where the initial topography was particularly flat and antecedent rivers where not deeply incised before the uplift.

Fig. 10. Last stage of the oblique convergence model. (a) Surface lithology (yellow for sediments; green for basement; brown for allochthonous basement; blue below sea level) and drainage (river thickness proportional to water discharge); (b) sediment thickness (contour interval is 1000 m) and transport rate (river thickness proportional to mass transport rate); (c) cross section along the transect located with a red line in panel (a). Note the location of the main axial river in the featheredge of the foreland basin.

Fig. 11. (a) Evolution of cumulative rock volume eroded in the oblique convergence model (dashed line) and sediment deposited on the initially subaerial foreland (bold line). Arrows indicate the periods of tectonic block movement; (b) rate of variation of the magnitudes in (a). Other parameters used for this model are listed in Table 3.
**DISCUSSION**

Numerical modelling and simulation help to understand whether the complexity of nature can be described as the addition of simple processes. In the context of the feedback's between processes depicted in Fig. 2, the numerical experiments shown in the previous sections provide a new insight on how flexure couples with the fluvial network to determine the surface transport in orogen/basin systems. This coupling is twofold: First, flexural vertical movements and the generation of accommodation space. In both cases, sediment accumulation is mostly restricted to the areas initially below sea level (see initial stage in Fig. 9).

A canonical type of foreland basins can be defined according to their drainage pattern: distal-drainage foreland basins are characterized by longitudinal transport through a main river flowing along the external (distal) edge of the basin. Within the approaches of the model designed here, a foreland basin is expected to belong to the distal-drainage type if (1) Tectonic loading of the lithosphere occurs only from one direction; (2) The dom-
iniant transport mechanism in the basin is the fluvial network; (3) The flexural forebulge has not been overflowed with sediments; and (4) surface transport in the system exceeds the tectonic growth of the orogen. This basin type is recognized in foreland basin systems such as the North-Alpine and South-Carpathian basins (Danube River), the Guadalquivir-Betic Basin (Guadalquivir River), or the Ganges-Himalayan system (Ganges River) (Fig. 3).

This distal drainage pattern is the result of interaction between river transport dynamics and lithosphere flexural response to mass redistribution: the expulsion of the main river towards the foreland (due to the unloading of sediment along the tributaries coming from the active margin) competes with the basin subsidence and with the barrier build in the foreland by flexural forebulge uplift. Therefore, the results provide a quantitative validation of the hypothesis by Burbank (1992) that drainage pattern in foreland basins is controlled mainly by the competition between sediment supply and flexural subsidence. The distal location of the main river is less probable during a syn-tectonic phase, because then flexure tilts the basin towards the orogen, whereas in the post-tectonic phase orogen erosion tilts the basin towards the foreland.

Most foreland basins not belonging to the distal drainage type fail some of the above conditions, such as those formed in multiply verging settings (tectonic loading from two or more directions), such as the Po (Apennines-Alps) and Ebro (Pyrenees-Iberian Chain) basins (e.g. Coney et al., 1996; Bertotti et al., 1998). The lack of distal drainage in other foreland basins can be related to variations in sediment supply/tectonic deformation ratio (Indo River; Burbank, 1992) or to processes not explicitly considered in this work such as the large-scale alteration of the drainage system by glaciations (Alberta Basin) or the relevance of non-fluvial transport (e.g. Zagros Basin).

In the models above, the cessation of orogenic crustal shortening determines the beginning of a transition period from an underfilled marine or lacustrine basin to a filled basin with continental deposition. This is related to the assumption that tectonic shortening occurs with a constant velocity and ends abruptly. A progressive slow down of the tectonic deformation could explain why this transition is sometimes observed during the late syn-orogenic period, as in the Swiss Molasse Basin (e.g. Sinclair & Allen, 1992; Schlunegger et al., 1997). Very high erosion rates can also overfill a basin during the syn-tectonic period.

The axial drainage towards the marginal sea is strongly facilitated by lateral variations of the initial topography along the orogen as those used in the oblique convergence model (Fig. 9). However, the results from the frontal convergence model (Fig. 6) show that axial drainage develops also in conditions of symmetric initial topography and synchronous tectonism. Tectonic loading can localize maximum flexural subsidence in front of the centre of the orogen (e.g. van Wees & Cloetingh, 1994), and induce the formation of large lakes being maintained tectonically for long periods and finding their outlet in the axial direction. Despite the two models assume synchronous orogen development, both predict axial drainage. Therefore, both the initial lateral variations of topography in the basin and the formation of lakes in local topographic minima are key processes developing axial drainage, without necessarily invoking diachronous tectonism as proposed by Johnson & Beaumont (1995). This explains why in fact most foreland basins, both in nature and in the computer simulations above, present axial drainage.

The feedback between transport capacity and the evolution of the topography is responsible for the complex, non-linear dynamics of basin evolution. Small perturbations on the initial topography and/or the tectonic kinematics used in the experiments lead to large changes in the river network, reflecting the non-linear character of river network self-organization (Braun & Sambridge, 1997). However, both crustal deformation and lithospheric flexure control and organise the large-scale drainage pattern, competing with its unpredictable intrinsic nature. This is reflected in the fact that the model described here predicts some general patterns that can be recognized in the field (axial, distal drainage and, possibly, forebulge drainage divide). The relevance of tectonics and flexure on the foreland fluvial network depends on the initial topography and the degree of organization of the drainage network. For example, well-developed (incised) antecedent river networks are hardly modified by the small vertical movements related to forebulge uplift.

**Interplay between flexure and fluvial transport in foreland basins**

The sediment volume accumulated in a foreland basin depends on the amount of tectonic shortening, the lithospheric rigidity and the transport capacity of rivers. The accepted idea that the formation of foreland basins implies a flexural behaviour of the lithosphere and that higher lithospheric rigidity implies larger sediment volume (e.g. Flemings & Jordan, 1989) is here revised through a closer approach to the surface processes. In some basins, most of the subsidence can be explained by isostatic compensation of the sediment load (e.g. Schlunegger et al., 1997), questioning the importance of the tectonic flexural subsidence. The results above show that some tectonic subsidence is necessary to trigger large deposition, because it provides initial space for the accumulation of sediments. As shown by models with null or low elastic thickness (Fig. 12a), the dynamics of fluvial transport (e.g. the unloading of sediments of rivers reaching the basin) are insufficient to accumulate important amounts of sediments in absence of accommodation space.

Surprisingly, $T_c$ values larger than 40 km also predict smaller basin volumes (Fig. 8d), in contrast with what diffusive transport models predict. Large $T_c$ values imply larger topographic contrast between orogen and basin, and thus larger erosion. Diffusive models deposit all eroded material next to the orogen predicting large basin volumes also for high $T_c$ values (Flemings & Jordan, 1989); Incor-
orating fluvial transport, the shallow, widespread subsidence predicted for large $T_c$ values is rapidly filled, and axial drainage drives the sediments far from the orogen/basin system, reducing the flexural subsidence and amplifying the erosional rebound. As tectonic deformation stops, the higher erosion rates for large $T_c$ values enhance the flexural rebound of the system (Heller et al., 1988) and rework part of the basin sediments. As a result, maximum sediment accumulation is predicted for basins developing on lithospheric plates with equivalent elastic thickness between 15 and 40 km (Fig. 8a). It is remarkable the good fit between this $T_c$ range and the $T_c$ values most frequently found in continental areas (Watts, 1992). This suggests that the values of equivalent elastic thickness derived for continents may be related not only to the mechanical properties of the lithosphere but also to their coupling with fluvial transport during the formation of foreland basins.

The onset of a constant-rate tectonic shortening induces a progressive increase of erosion and deposition rates towards a dynamic equilibrium between tectonism and erosion (Fig. 8a). Thus, maximum denudation rates are reached at the end of the tectonic phase, in agreement with the results by Kooi & Beaumont (1996), and then a transition towards a new steady state (penepalzation) begins. However, the results show here that the cessation of tectonism does not always coincide with the maximum sediment volume of the basin, which can be significantly delayed up to $\sim$10 Myr depending on how starved the basin reached the post-tectonic phase. Another striking result is that an increase in river transport capacity by, for example, higher precipitation rates implies lower basin volumes (compare Figs 8b and 8c). This is because associated erosional unload in the orogen is larger than the load deposited in the basin, since most sediments are transported out of the basin as soon as this is filled.

The application of the model proposed in this work to understand the mass balance of specific orogen/basin systems is a challenge requiring the integration of a diversity of data: palaeoclimate, hydrogeology, palaeogeography, structural geology, sedimentary basin analysis and lithosphere tectonophysics. As return, this technique can provide a quantitative link between the increasing knowledge on the 3D sediment balance between orogen and foreland basin (e.g. Schlunegger et al., 2001; Willet & Brandon, 2002) and the mechanisms controlling the drainage patterns (Burbank, 1992) and the landscape evolution of foreland basins.

CONCLUSIONS

Based on the numerical experiments undertaken in this work and subject to the previous discussion, the following conclusions can be made:

1. The bending of the lithosphere under the orogenic load controls the drainage pattern of foreland basins. The axial and distal drainage pattern recognized in many foreland basins is a result of the competition between flexural tilting of the basin and sediment supply from the orogen.

2. Lateral variations of the initial topography and/or diachronous evolution of the orogen both facilitate axial drainage, but are not essential for it. In a perfectly symmetric orogenic evolution, the installation of a longitudinal drainage is also probable due to the formation of large lakes along the basin that find their outlet in the axial direction.

3. The axial fluvial transport in foreland basins has an important influence on the mass balance between erosion and sedimentation, since erosion products can be driven far from the orogen/basin system. This has in turn an effect on the vertical flexural movements and the generation of accommodation space, leading to maximum basin sediment volumes at intermediate values of lithospheric elastic thickness, between 15 and 40 km.

ACKNOWLEDGEMENTS

Manel Fernández and Montserrat Torne (Earth Sciences Institute Jaume Almera, CSIC, Barcelona) promoted and facilitated the realisation of this work. I want to acknowledge the particularly wise, constructive and challenging revisions of the original manuscript by Fritz Schlunegger and Sean Willet. Rich discussions with Prof Christopher Beaumont and Philippe Fullsack were useful during the model development. Prof Sierd Cloetingh and many others have encouraged the writing of the manuscript. A number of Spanish, Dutch and European public institutions provided financial support and infrastructure.

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Manuscript received 2 May 2001; Manuscript accepted 28 March 2002.
