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Long-term evolution of tectonic lakes: Climatic controls on the development of internally drained basins

Daniel Garcia-Castellanos†
Netherlands Centre for Integrated Solid Earth Science, Faculty of Earth and Life Sciences, Vrije Universiteit Amsterdam, De Boelelaan 1085, 1081HV Amsterdam, Netherlands, and *Institute of Earth Sciences "Jaume Almera" (CSIC), Solé i Sabarís s/n, 08028 Barcelona, Spain*

ABSTRACT

A quantitative model is proposed for the long-term evolution of lakes and internally drained basins resulting from tectonic vertical motions, sediment infill, outlet erosion, and climatic regime. The model accounts for the formation of a water body in the topographic basin created by tectonic uplift across a river, where incision capability is calculated using a stream power-law. The model also addresses the notion that, after cessation of tectonic forcing, lakes are transitory phenomena over geological time scales. High uplift rates across an antecedent river, in combination with low upstream precipitation, result in river defeat and lake formation. In addition to geometrical, lithological, and tectonic parameters, the evaporation rate at the lake surface is revealed as a key factor triggering drainage closure (endorheism) and significant lake life extension by preventing outlet erosion. Post-tectonic lake extinction is ensured by sediment overfill and/or outlet erosion. Once uplift comes to an end and drainage reopens (lake capture), shallow lakes at high altitudes undergo a faster reintegration into the drainage network and extinction. Vertical isostatic movements of the lithosphere significantly delay this process in lakes larger than 50–200 km. The development of an internally drained basin out of an open lacustrine basin requires that uplift across the outlet persists until the lake is large enough to evaporate all collected water. The evolution of tectonic lakes has, therefore, similar dependency on geometrical constraints (initial relief, length of the river, hypsometry of the catchment), lithological parameters (rock erodibility), tectonics (uplift rate, duration, and its spatial distribution), and climate (precipitation and evaporation rates).

Keywords: endorheic, surface processes, landscape evolution, lithospheric flexure, numerical model.

INTRODUCTION

Lakes develop in local topographic minima formed by processes as diverse as crustal tectonic deformation, volcanism, glacial excavation, or meteoritic impacts. On geological time scales, lakes are ephemeral systems that, once the forcing mechanism ceases, disappear by outlet erosion and/or sediment overfill (e.g., Ollier, 1981). Most lakes on Earth are related to the last glaciation and are therefore relatively recent. In contrast, most long-living lakes lasting beyond glacial-interglacial cycles are maintained by very active tectonic processes, such as continental rifting (e.g., Lake Baikal [Williams et al., 1997], 31,500 km^2 in surface and Pliocene-Quaternary in age) or continental compression (e.g., Lake Titicaca in the Andean Altiplano, 8,000 km^2, at

†E-mail: danielgc@ija.csic. es.

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least Pliocene in age) interfering with an existing river network. Independently of the tectonic mechanism generating the tectonic barrier, it is often interpreted that the formation of a tectonic lake is coeval with fast vertical movements (Fig. 1A) and that its extinction coincides with slower or negligible tectonism and/or increased erosion power across the lake outlet. An example for this is Lake Bungunnia, a 33,000 km^2 water body existing in southern Australia between 2.5 and 0.7 Ma, thought to be a result of the rise of a tectonic dam (Stephenson, 1986) between the Murray Basin and the ocean. Dry climate (low precipitation/evaporation ratios), sometimes in combination with active tectonism, can neutralize the lake outlet, closing its drainage (endorheic basins such as the present Andean Altiplano, the Dead Sea, the Caspian Sea, the Aral Sea) and presumably contributing to the extension of the life of the lacustrine system.

A quantitative understanding of the long-term evolution of lakes and internally drained basins is crucial because these systems are climate-sensitive discontinuities in the sediment flow from orogens to basins, and therefore their sedimentary infill constitutes a record of the climatic and tectonic evolution (e.g., Tiercelin, 2002; Yan et al., 2002). Intramountain basins, often presenting long lacustrine periods (e.g., Burbank, 1983), are conspicuous examples of interplay between climate and tectonics. In an open lake, the amount of output water discharge is reduced by evaporation from its surface (Fig. 1A). If evaporation becomes larger than the collected precipitation, the lake becomes the center of a closed (endorheic) drainage basin, under which circumstances the lake's level is climatically controlled (e.g., Lake Issyk-Kul; De Batist et al., 2002). An extreme example of endorheism is the Dead Sea, where evaporation reaches values of 1600 mm/yr, whereas the mean water runoff collected by the Jordan River (mean discharge of 17.1 m^3/s) is only 0.013 mm/yr. In Lake Titicaca, the evaporation rate of 1100 mm/yr in the lake's surface nearly matches the water it collects from the Altiplano, where precipitation rates range between 600 and 800 mm/yr. In the Minchin paleolake (Pleistocene highstand of the intramountain Altiplano Basin [Fornari et al., 2001], with a water surface larger than 40,000 km^2), the mean precipitation rate must have been higher than 300 mm/yr to compensate an evaporation rate similar to the present one. In such cases, due to the absence of river incision along an outlet, the lake's life expectancy is likely to increase. In closed drainage basins, either a climate change to wetter climatic conditions, or the sediment overfill of the lake (both inducing lake level rise) or a basin capture undertaken by an adjacent river can trigger drainage opening (Garcia-Castellanos et al., 2003), abruptly activating basin-wide incision and

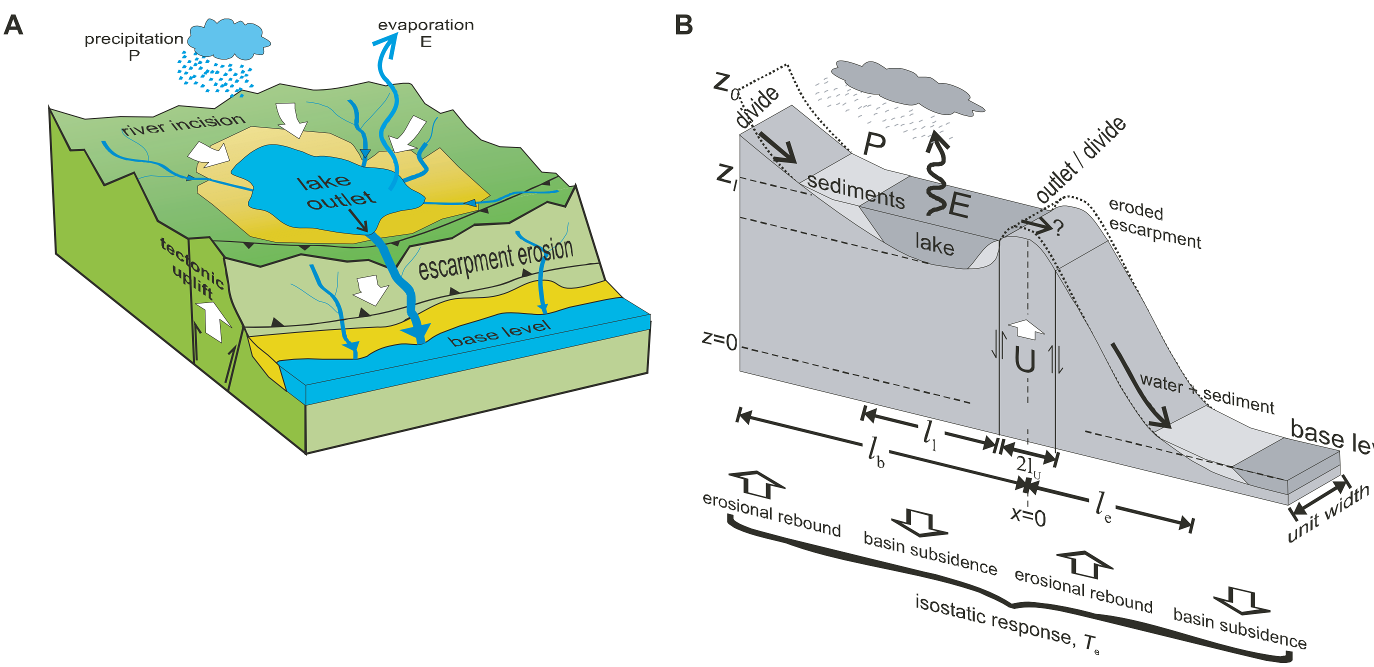


Figure 1. (A) Cartoon showing the main processes involved in the evolution of tectonic lakes, including the growth of a tectonic topographic barrier, river transport of sediments from the catchment toward the lake, erosion and river incision of the barrier, and the hydrological balance between precipitation, lake evaporation, and outlet discharge. (B) One-dimensional (1D, along-stream) approach and main tunable parameters of the cross-section model proposed here: P—collected precipitation (runoff, in mm/yr); E—evaporation rate (mm/yr); U—uplift rate (mm/yr); l_u—length of the uplifted area; l_b—basin length; l_e—escarpment length. As a result of the numerical simulation, lake altitude z_l, lake length along the modeling plane l_l, and other properties are predicted through time; thin arrows indicate the flow of water. Low P/E ratios can trigger drainage closure and endorheism, inhibiting water flow and erosion at the uplifting barrier. Large white arrows indicate fluvial and tectonic mass transport and vertical tectonic movements.

lake extinction, as reported in a diversity of geological scenarios including the Pleistocene Lake Bonneville (O'Connor, 1993) and the Ma'adim Vallis in Mars (Irwin et al., 2002). Again, though these processes are often cited in the literature as governing the extinction of lacustrine basins, little is known about the relative importance of the parameters involved, and a quantitative study of the process interplay between hydrology, erosion, and tectonics is needed.

Studies of river incision and sediment transport have provided quantitative models of landscape evolution driven by surface water flow (e.g., Howard, 1994). These models have been useful to quantitatively understand the dynamics of tectonic defeat of antecedent rivers (Kühni and Pfiffner, 2001; van der Beek et al., 2002; Sobel et al., 2003). These studies conclude that the main controlling factors are the mean precipitation in the catchment, the rate of tectonic uplift across the river path, and the rock strength of the uplifting barrier. However, these models do not explicitly account for the formation of a lake in front of the uplifted area and its effects on the water balance between collected runoff and evaporation. This is of key relevance for internally drained basins, because by definition the transition from an intramountain basin to an internally drained basin requires the evaporation at the lake surface to compensate the collected water runoff. Even before drainage closure is reached by tectonic uplift, evaporation reduces water discharge across the uplifting area, limiting the erosion rates and presumably accelerating the drainage closure process.

Coupling well-established quantitative approaches to both tectonic and surface processes, this paper focuses on the interplay between these two scales in controlling the long-term evolution of lakes. A cross-sectional model of lake evolution is designed to account for four processes: tectonic uplift of a structural barrier, hydrological balance, surface sediment transport, and flexural isostasy. The model permits parameterization of the potential formation of lakes and closed (internally drained) basins, their sediment infill, and their extinction as a function of the timing of tectonic uplift.

FUNDAMENTALS AND FORMULATION

The scenario of a river undergoing tectonic uplift along part of its path is addressed via a one-dimensional (1D, along-stream cross section) numerical model of river incision and transport. Although a 1D approach might seem an oversimplification, it will be later shown that other assumptions inherent to the adopted transport model impede a direct quantitative comparison with geological scenarios. Thus, a 1D model is chosen as the easiest way to simulate the process interplay while allowing an easy reproducibility of the results. Note also that using a 2D (planform) model would introduce infinite further arbitrary choices for the initial topographic setting. The present approach is similar to the along-stream approach used by Sobel et al. (2003; based on Howard, 1994, among others) to calculate the defeat of rivers by tectonic uplift, but here lakes are explicitly considered and assumed to have the same width (in the direction perpendicular to the modeling plane) as the hydrographic basin.

The lake's history is predicted as a result of the competition between tectonic uplift versus river incision and transport. Due to the 1D nature of the model, hereafter the length of a feature measured along the modeling plane will be assumed to be equivalent to the area of that feature in nature. The origin of the along-stream coordinate x is chosen to be at the center of the uplifting barrier, which has a half-length l_u and is assumed to be invariable along the strike (Fig. 1B). Tectonic uplift is incorporated to the model as a rectangle function $U(x) = U$ for $x = [-l_u, l_u]$ and $U(x) = 0$ elsewhere, and is applied from the initial time $t = 0$ until $t = t_u$ (thus, the period after t_u is referred to as post-tectonic). In accord with the cross-section nature of the model, the lake of length l_l forming in front of this uplifting barrier has a unit width along strike. The adopted initial topography (Figs. 1B and 2A) corresponds to an equilibrium river profile (i.e., one with erosion rate constant along the river) descending from an altitude z_0 at $x = -l_b$ (left boundary of the model) to zero at $x = l_e$. At $x > l_e$, a basin is defined with a constant water level at $z = 0$, which plays the role of a geomorphological base level.

Erosion is modeled via a combination of fluvial incision and diffusive rock flow. These processes are usually represented via a stream power law and a diffusive transport law, respectively (Howard, 1994; Tucker and Slingerland, 1997; Whipple and Tucker, 1999), giving the erosion rate e at any given location x of the topographic profile $z(x)$ as:

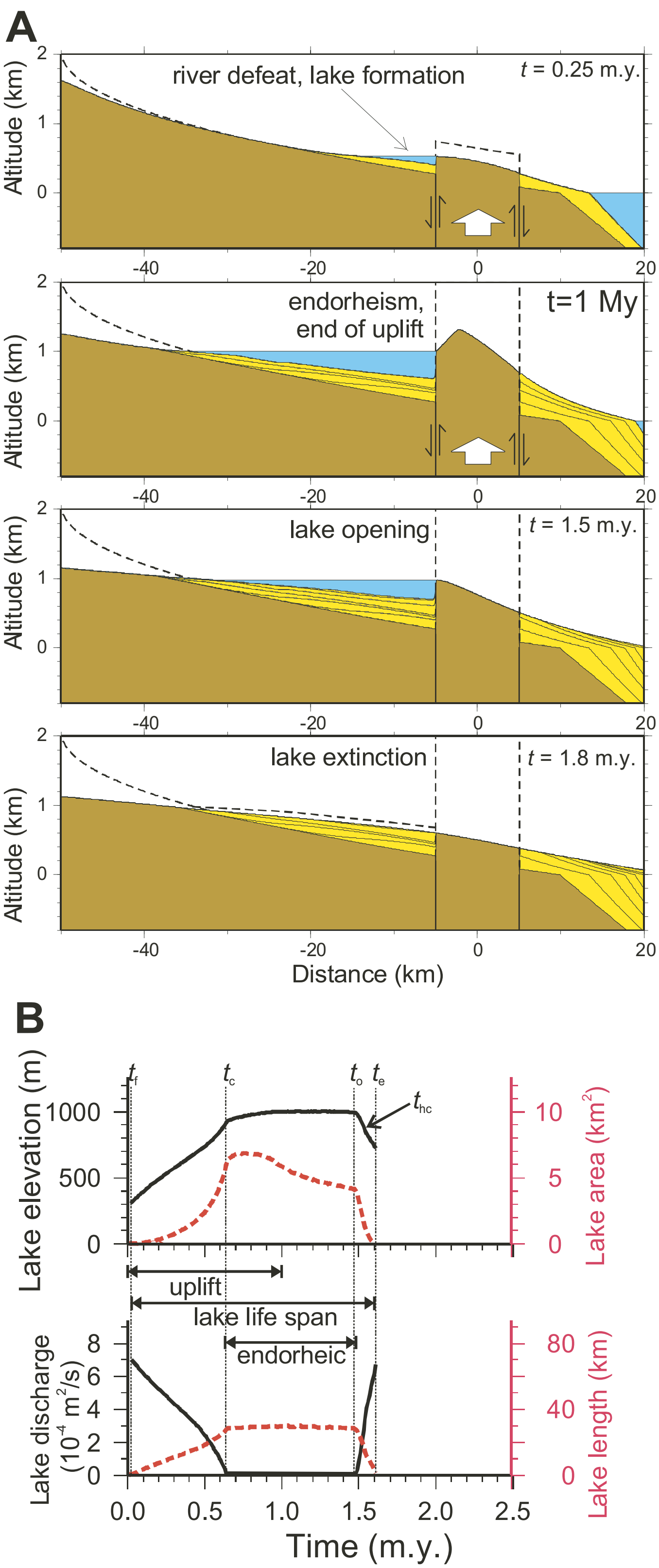
$$e(x) = -Kd\frac{\partial^2 z}{\partial x^2} + K_f Q_w{}^m \left|\frac{\partial z}{\partial x}\right|^n, \qquad (1)$$

where K_d is the constant of diffusive transport, K_f is the constant of fluvial incision or erodibility, and Q_w is the water discharge. Note the different units used here for K_f (Table 1) with respect to the previous references. The exponents m and n determine the relative importance of water discharge against slope in driving fluvial incision, and a value of $n/m \approx 2$ is generally accepted based on field data (e.g., van der Beek and Bishop, 2003). The values $m = 1/3$ and $n = 2/3$ are adopted here, corresponding to the assumption that incision rate is proportional to basal shear stress in the flowing water (Howard, 1994; Whipple and Tucker, 1999). The dependence of K_f on lithology and other parameters is not well understood, and the values reported in the literature are highly variable. Besides that, this model overlooks the role of the sediment load and its granulometry in defining the incision rates. These uncertainties indicate that the results from the model cannot be directly compared with a particular geological scenario without performing previously a detailed calibration. Finally, note that the first term in equation 1 is incorporated to produce erosion in areas where slope is zero, such as the outlet saddle of a lake. Rather than representing hillslope processes, it is here intended to capture any other nonfluvial erosion and transport mechanism, such as water-flow shear stresses, vegetation, frost, etc.

In the absence of significant underground water flow, the water discharge leaving a lake is determined by the water balance

TABLE 1. INPUT PARAMETER VALUES OF THE REFERENCE MODEL SHOWN IN FIGURE 2

Parameter	Variable	Value
Tectonics		
Uplift rate	U	2 mm/yr
Uplift duration	t_u	1 m.y.
Uplift area half-length	l_u	5 km
Elastic thickness (km)	T_e	∞
Climate		
Precipitation rate	P	500 mm/yr
Evaporation rate	E	800 mm/yr
Lithology		
Erodibility	K_f	
bedrock		1.3 10^{-8} m/s $(m^2/s)^{1/3}$
sediment		1.3 10^{-7} m/s $(m^2/s)^{1/3}$
Diffusive transp. coef	K_d	0.01 m^2/yr
Geometry		
Max. initial topography	z_0	2000 m
Basin length	l_b	50 km
Escarpment length	l_e	10 km



between precipitation and evaporation, which for a steady-state water flow can be written as

$$Q_w(x = 0) = P \cdot l_b - E \cdot l_l, \tag{2}$$

where $Q_w(x = 0)$ is the water discharge at the outlet, P is the mean precipitation rate in the lake's catchment basin of length l_b (measured along the modeling plane and assumed to be invariable along the strike), E is the evaporation rate at the lake's surface, and l_l is the length of the lake (representing its area in real three-dimensional nature). Note that here the term precipitation is used to designate the component of precipitated water that is collected as surface runoff, this to avoid addressing local infiltration and evapotranspiration processes that are out of the scope of this paper. By definition, for a closed lake $Q_w(x = 0) = 0$ and therefore $P/E = l_l/l_b < 0$, whereas for an open lake (in which the lake's water level reaches the outlet), the output discharge is $P \cdot l_b - E \cdot l_l$, which

Figure 2. (A) Time evolution of the reference model (Table 1). Eroded rock is indicated with a dashed line. Sediment horizons are every 0.25 m.y.; vertical exaggeration ×8. (B) Evolution of lake altitude, lake area across the modeling plane (representing volume in the 3D nature, dashed line in top panel), lake length l_l (representing surface, dashed line in bottom panel), and outlet water discharge of the lake. A lake forms in this example between $t_f = 0.02$ m.y. until $t_e = 1.62$ m.y. Lake length and elevation increase until evaporation on its surface equals the precipitation collected in the lake. At that stage (time of closure, t_c), the output water discharge becomes zero and the lake becomes endorheic. After the cessation of uplift at $t_u = 1$ m.y., the altitude of the drainage divide between the lake and the ocean side decreases until it reaches the lake level (time of drainage opening t_o). The lake extinguishes at t_e, both by sediment infill and outlet erosion. The parameter t_{hc} is the time of half capture (time at which closing the drainage again would require a reduction in the precipitation/evaporation ratio of a factor 2, see text).

produces outlet incision leading to the integration of the lake in the drainage network (drainage capture).

The area of a lake is related to its water level, z_l, through the hypsometric curve of its floor topography. Within the adopted 1D approach, this hypsometric curve is expressed in terms of lake length $l_l(z)$ (representing lake surface in nature) below an altitude z, and coincides with the lake bathymetric profile. At the time of drainage opening, by definition, $l_l(z_{min}) = 0$ and $l_l(z_{max}) = l_b \cdot P/E$, where z_{min} corresponds to the minimum elevation of the lake floor, and z_{max} corresponds to the outlet level at the time of opening. During opening, the excess water leaving the lake (i.e., the water that would have been evaporated in the lake portion lying now above z_l) is related to the lake level z_l, and the water discharge along the outlet river follows the equation

$$Q_w(x) = Pl_b - El_l(z_l) + K_H x^h, \tag{3}$$

where x is measured from the outlet in opposite direction to the lake (i.e., toward the escarpment) and the second term corresponds to Hack's Law of river water discharge. For simplicity, $h = 1$ is assumed (which implies that the planform geometry of the catchment, including the lake, is rectangular), and then K_H coincides with the precipitation P, which is assumed to be constant over the entire region. Measured values of h often range between 1.55 and 1.65 in the higher parts of a catchment (e.g., Rodriguez-Iturbe and Rinaldo, 1997), but are <1 in the lower parts. For intramountain lakes, our experiments overestimate the amount of water reaching the lake relative to that leaving it. For the purpose of the present work, it is sufficient to keep in mind that higher values of the Hack's Law exponent would facilitate the drainage closure of the basin, not affecting the qualitative role played by the mean precipitation P and other parameters. From equations 1 and 3, the changes in elevation along the outlet river profile $z(x)$ are:

$$\begin{aligned}\frac{dz}{dt} &= U(x) - e(x) = U_T(x) + U_I(x) + K_d \frac{\partial^2 z}{\partial x^2} \\ &- K_f (\mathrm{P\,I_b} - \mathrm{E\,I_1(z_1)} + P \cdot x^h)^m \left|\frac{\partial z}{\partial x}\right|^n\end{aligned} \tag{4}$$

where $U(x)$ is the rock uplift rate resulting from tectonic deformation $U_T(x)$ and flexural isostatic readjustments $U_I(x)$. Equation 4 reflects that the reduction of $l_l(z_l)$ as the outlet is eroded produces a feedback effect that initially accelerates the capture process. Note that slope is zero by definition in the outlet, and therefore outlet erosion corresponds only to the third (diffusive) term in the right side of equation 4. However, since curvature in the outlet area is proportional to the slope imposed by fluvial incision in the escarpment (Fig. 1), erosion rate of the outlet is also indirectly controlled by water discharge. This implies that equation 4 cannot be reduced to a simple version restricted to the outlet point, and numerical solutions of the entire domain are required. On the other hand, reasonable values of K_d between 10^{-3} and 10^{-1} m^2/yr (e.g., Flemings and Jordan, 1989; Braun et al., 2001) have a negligible effect on the numerically predicted lake evolutions, because the discretization of topography applied in the numerical method implies nonzero slopes also in an outlet or a drainage divide, allowing for the erosion of such points in the same way as diffusion does. In conclusion, though the diffusive term is analytically required, it has no influence on the numerical results.

Equations 1 and 4 are solved numerically with an improved version of the code *tAo* (Garcia-Castellanos et al., 1997, 2002; http://cuba.ija.csic.es/~danielgc/lakes/), which uses the finite difference technique with a regular grid (dx = 100 m in this work), in three steps: (1) determination of the drainage flow following a maximum slope criteria and identifying the local topographic minima; (2) determination of water discharges in every location, and the extension of lakes accounting for evaporation; and (3) determination of the amount of material eroded/deposited at each location. Iterating these at time steps of 20 yr yields to the topographic and drainage evolutions described later in this paper.

To check the possible influence of isostasy on lake evolution, surface mass transport and changes in water volume in the lake are isostatically compensated via a 1D thin-plate flexural model with laterally constant elastic thickness T_e. Large T_e values imply a more rigid behavior of the lithosphere under surface mass transport, resulting in longer wavelength distribution of vertical movements $U_I(x)$ in equation 4 (e.g., Turcotte and Schubert, 2002). In contrast, small T_e values imply that erosional unloading and sediment loading are locally compensated by vertical rebound and subsidence, respectively.

LAKE FORMATION BY TECTONIC UPLIFT ACROSS A RIVER

A reference model setting is designed to check the role of tectonic uplift, climate, and rock strength on the initiation and evolution of lakes (Fig. 2). For the sake of simplicity, precipitation rate P, lake evaporation rate E, and uplift rate $U(x)$, are set constant through time. Uplift $U(x)$ stops at t_u = 2 m.y., and is localized along a 10-km-long area (l_u = 5 km) centered at $x = 0$, next to the base level. Other parameters are listed in Table 1.

For fast tectonic uplift rates and relatively low precipitation (P) and initial topographic gradients (low z_0), incision along the river is defeated by uplift, and a local topographic minimum develops in front of the uplifting area. In the model, this coincides with the onset of river aggradation in the upstream side of the tectonic barrier. If river aggradation cannot pace with uplift (e.g., if sediment yield is low due to low topography gradient), then a lake develops. For the parameter values used for the reference model (Table 1; model evolution in Fig. 2), the lake formation time is t_f = 0.02 m.y. Double precipitation rate (1000 mm/yr) results in more efficient river incision and later lake formation (t_f = 0.06 m.y.). A slower uplift rate of U_T = 0.5 mm/yr (compared to the 2.0 mm/yr adopted for the reference model) results also in a longer alluvial aggradation period, delaying the lake formation until t = 0.21 m.y. Other parameters, such as low bedrock erodibility and/or low transport coefficient, facilitate river defeat, similar to the predictions by Sobel et al. (2003).

Water accumulates in the new basin forming a lake with the outlet located in the topographic maximum of the uplifted area. Whereas in Sobel et al. (2003) the initiation of internal drainage is considered coincident with the defeat of the river, in the present model river defeat implies only the formation of a lake, and internal drainage only develops if the lake surface becomes large enough to evaporate the water discharge collected by the lake. Before this condition is accomplished, the lake remains open, and as the outlet is uplifted and incised by the outlet river, so changes the lake level. The difference between river defeat and internal drainage is quantitatively important. The minimum uplift rate required to defeat the river (keeping all other parameter values as in Table 1) is 0.35 mm/yr, whereas in order to create a closed drainage basin, the uplift rate must be at least 1.12 mm/yr. Similarly, the topographic difference between the mountain-side boundary of the model (x =) and the base level also influences the timing of lake formation and drainage closure, since higher initial slopes induce a more efficient erosion of the uplifting barrier. Whereas z_0 > 3810 m is required to inhibit river defeat (all other parameters being as in Table 1), the development of internal drainage is impeded for z_0 > 2780 m.

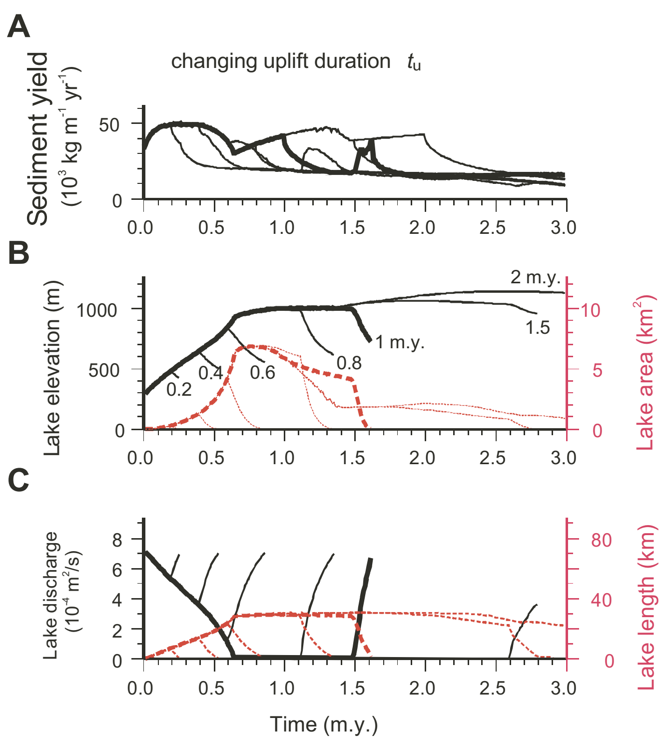


Figure 3. Evolution of the reference model in Figure 2 and Table 1 (bold lines) and a set of models differing only in uplift duration t_u (thin lines; labels indicate t_u). (A) Sediment yield toward the lake and the base level (identical to the erosion rate in the whole model domain); (B) lake level and lake volume; (C) output water discharge and lake area. High uplift rates and/or high uplift durations increase the duration of the lake, particularly if drainage closure (endorheism) is attained. Other legend as in Figure 2B.

If the increasing lake length reaches the critical value at which the potential evaporation at its surface becomes larger than the water collected (equation 2), then the lake becomes closed (t_c = 0.62 m.y. in the reference model in Fig. 2). This situation is similar to that of the present Lake Issyk-Kul (De Batist et al., 2002), which has gradually become endorheic since the late Pleistocene. Another example is the closure of the Ebro Basin (NE Iberia) 35 m.y. ago as a result of tectonic shortening in the Pyrenees and other surrounding mountain belts (Garcia-Castellanos et al., 2003). In the model, the occurrence of closure depends on the uplift rate, U, and its duration, t_u, on the climatic ration P/E and on the ratio between l_b and l_l. As expected, long uplift periods (large t_u) promote long closed-drainage periods (Fig. 3), and low uplift rates delay the time of closure t_c (Fig. 4).

Once the lake becomes closed, the lake level (altitude of the lake surface) stops increasing, because it becomes controlled by the hydrological balance, rather than by the altitude of the outlet. The small increase in lake level during this period (Figs. 3B and 4A) is related only to the progressive sedimentation near the lake's shore (alluvial aggradation above the lake level; lacustrine progradation within the lake): the sediment delivered to the lake displaces its shore, so the lake's altitude must increase to preserve a length sufficient to evaporate the collected water (equation 2). The lake keeps a nearly constant length (l_l) as long as it remains closed, because the collected water remains nearly invariable (the internal catchment is only slightly reduced by divide retreat in the topographic barrier). Drainage closure promotes additional sediment supply to the lake basin coming from the uplifted area,

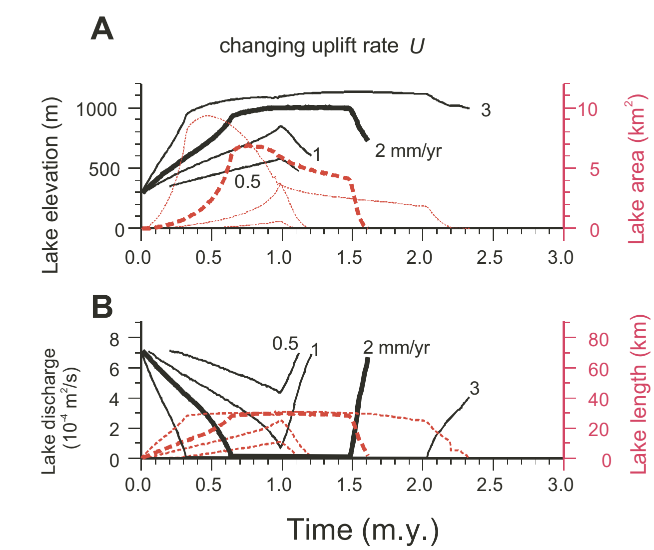
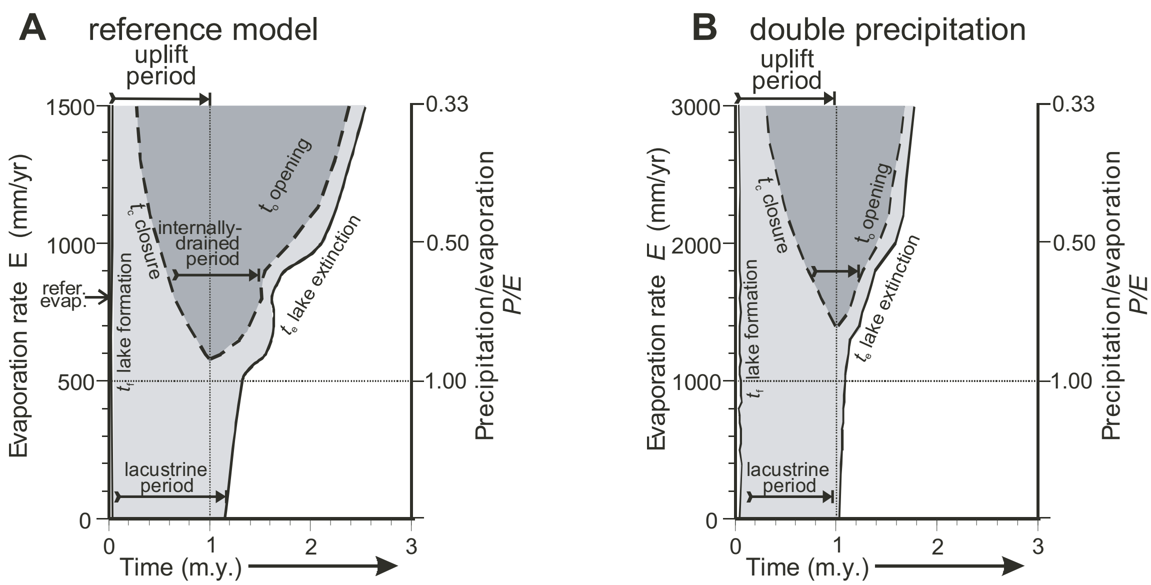


Figure 4. (A) Evolution of lake elevation (bold line) and lake length (dashed line) along the modeling plane; and (B) evolution of lake area (dashed line) and output discharge (bold line) of the reference model shown in Figure 2 and Table 1 (bold lines), and a set of models differing only in uplift rate U (thin lines; labels indicate U in mm/yr). Note the abrupt increase in lake life if the uplift rate is enough to close the drainage (i.e., if evaporation reduces output discharge to zero).

shifting the highest deposition rates toward the outlet side of the lake (Fig. 2A, t = 1 m.y.). Nevertheless, the total erosion rate occurring in the whole model domain reaches a minimum at t_c as a result of the reduced water flow at that time (Fig. 3A).

According to the model predictions, lake evaporation exerts a key control on lake evolution and on the development of internally drained basins. In order to become a closed drainage basin, the lake must attain a sufficient area (length in the present 1D model) so that evaporation in its surface equals the collected water. The dependency of the timing of lake development on evaporation is shown in Figure 5. The time of lake formation is independent of the evaporation rate (t_f = 0.02 m.y.), because evaporation only takes place at the lake surface (evapotranspiration in the rest of the model is not implemented in the model for simplicity). Lake closure (endorheism) occurs only for evaporation rates higher than the assumed precipitation (Fig. 5; Table 1), and it significantly increases the duration of the lake t_e–t_f.

As tectonic uplift comes to an end, the barrier separating the closed lake from the base level is eroded until its maximum topography becomes lower than the lake's level. This corresponds to the time of drainage opening t_o (1.48 m.y. in the reference model), initiating the process of lake capture. During capture, maximum erosion rates in the outlet are reached as a result of the increased water discharge in that area (Fig. 3A). Erosion of the barrier decreases the lake elevation z_l and length l_l, inducing a progressive increase in the water discharge delivered to the escarpment. This increase in water discharge induces an increase of erosion at the escarpment and thus a faster lake-level fall. As



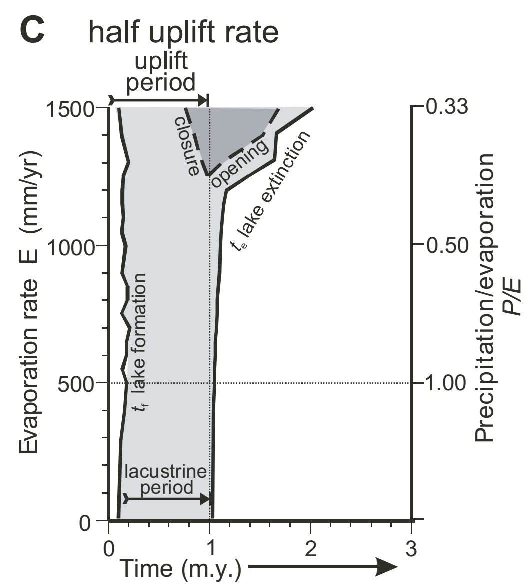


Figure 5. Effect of the climatic parameters P, E (precipitation and evaporation rate) on the time of lake formation t_f, time of drainage closure t_c, time of opening t_o, and time of lake extinction t_e. The light shade indicates the period during which a lake is formed; the dark shade indicates endorheism (closed drainage), lasting from t_c to t_o. (A) Results of the reference model for different evaporation rates (note that for the reference model in Figure 2, a precipitation rate of P = 500 mm/yr was used); (B) same as A for double the precipitation (1000 mm/yr); (C) same as A for half uplift rate (1 mm/yr). The time of formation (river defeat) is not dependent on E, but on precipitation (t_f = 0.2 m.y. for P = 500 mm/yr; t_f = 0.6 m.y. for P = 1000 mm/yr) and uplift rate. For values of $E > P$, the lake can undergo an endorheic period, which duration is proportional to E. This closed-drainage period significantly prolongs the life of the lake relative to lower E values. Though the general trends are similar, the different timings calculated in A and B indicate that the timing of the lake is dependent not only on the ratio P/E, but also on the absolute precipitation itself. Higher precipitation rates and lower uplift rates shorten the life of the lake by increasing erosion versus uplift in the outlet.

the capture of the entire lake progresses, this feedback effect is counteracted by the decreasing topographic difference between lake level and base level. The lake's bathymetry attained at the time of drainage opening also influences the changes in outlet erosion (equation 4). Closed lakes undergoing sediment overfilling disappear shortly after capture, because their floor is nearly flat and easy to integrate into the drainage network. The effect of these parameters in the post-tectonic lake opening and extinction deserves attention in a separate section.

OPENING AND EXTINCTION OF A LAKE

To show more conspicuously the effect of each parameter on the acceleration of the opening, a measure is necessary of the time at which closing the lake again (e.g., by a climatic change to dry conditions) becomes improbable. The time of half capture, t_{hc}, is here defined as the time at which the lake is reduced by a factor 2 in length, as a result of the ongoing capture. This is equivalent to define t_{hc} as the time at which a reduction by a factor 2 of the P/E ratio is required to close again the drainage (i.e., to decrease the lake level below the outlet level). A time of half capture (t_{hc}) closely after opening (t_o) implies that the drainage opening soon becomes irreversible because water discharge and incision rate in the outlet have quickly increased. Both t_{hc} and the time required for lake extinction t_e increase for large values of depth and length of the lake or for low precipitation values. Note that by definition, $t_o < t_{hc} < t_e$. A t_{hc} value closer to t_o than to t_e indicates that capture is faster at the beginning than at the end. In particular, lake extinction will not occur if the bottom of the lake is below the base level. Also, t_{hc} and t_e decrease nearly linearly with the ratio z_l/l_e (lake elevation divided by escarpment length). The model predicts that maximum incision rates in the outlet area and maximum sediment delivery at the base level occur (Fig. 3A) between t_{hc} and t_e. The time of lake extinction t_e is proportional to the evaporation rate, since endorheism impedes water discharge and erosion in the outlet.

The velocity of lake capture and its reintegration in the fluvial network is also strongly dependent on the ratio z_l/l_e (Fig. 6A). Lakes slightly above the base level undergo very little incision in the outlet, due to the small slope along the outlet river, and extinguish very slowly, mostly by sediment overfilling. High-altitude lakes, in contrast, disappear soon after opening, depending on the other geometrical parameters (e.g., slower if lake depth and length are large). For fixed values of lake depth and length at the time of opening, the lake hypsometry inherited at that stage has two opposing effects on the timing of capture. On one side, it controls the reduction in lake surface, as erosion in the outlet reduces its level. At t_o, a high lake-floor concavity (wide and flat bottom in the center of the lake and steep floor near the shore) implies a smaller reduction in lake length for a given outlet incision. In turn, this implies a smaller increase in water discharge and incision at the outlet, and therefore a slower capture. On the other hand, a high floor concavity also implies smaller distance between the lake's bottom and the base level, requiring shorter time to propagate erosion into the lake's center. The model results indicate that this latter effect dominates over the first and that lakes with high-concavity bathymetry suffer slower captures. Overall, however, the effect of the hypsometric concavity on the lake capture is secondary relative to the effects of lake altitude, evaporation, precipitation, and tectonic uplift, described above. Note also that the time of half capture, t_{hc} (Fig. 6), is always similar or smaller than half of t_e–t_o, showing that the first half of the capture moves at similar speed or faster than the second.

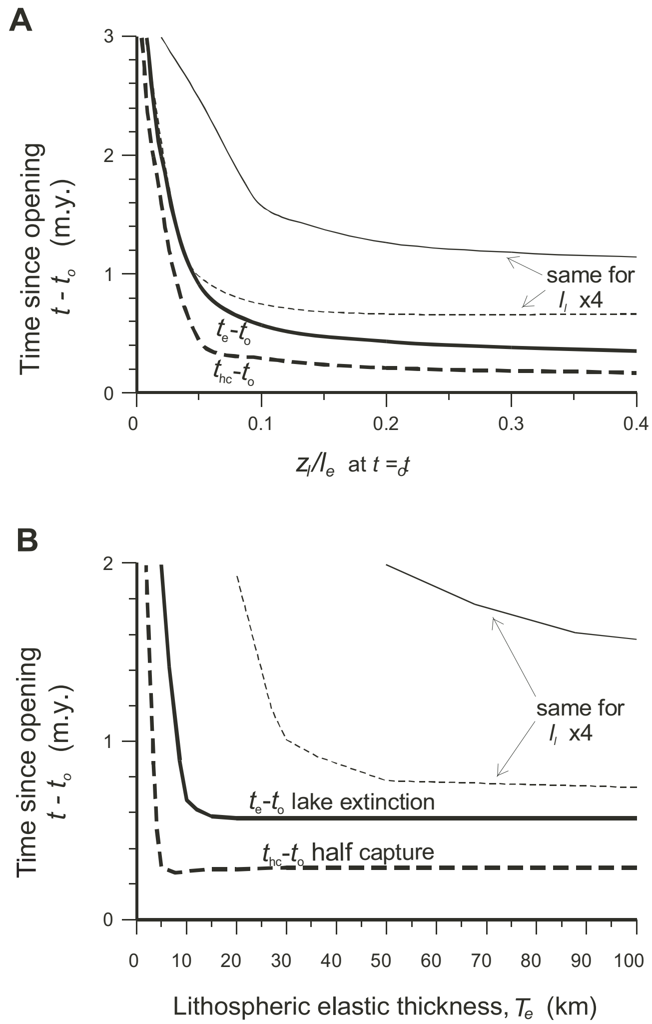


Figure 6. Timing of lake extinction after drainage opening. (A) Effect of lake elevation (normalized with the escarpment length l_e) on the time of half capture (dashed line) and time of extinction (plain line), both measured from the time of opening (t_{hc}–t_o and t_e–t_o, respectively). Thick lines show the results for a lake length of l_l = 50 km; thin lines are calculated for l_l = 200 km. Lakes elevated high above the base level and/or separated from the base level by a high-slope topographic barrier undergo a faster post-tectonic extinction. (B) Same as A, but varying the lithospheric elastic thickness T_e. Large T_e values reduce the isostatic rebound in the lake outlet in response to erosion, accelerating lake extinction.

Because the formation and closure of the lake depends strongly on the uplift rate and initial slope, whereas its opening and extinction depend rather on the accumulated uplift, it is not possible to find a simple parameterization of the lake's life based on the results of the model. Figure 7 shows an attempt to parameterize the lake history as a function of the adimensional quotient between the total uplift $U \cdot t_u$ and the maximum initial topography z_0. Note that the reference model was arbitrarily chosen so that $U \cdot t_u / z_0 = 1$ (Table 1). Comparison between Figures 7A and 7B shows that the regions of the parameter space where the lake forms and closes are similar, but significant differences remain, particularly for low values of $U \cdot t_u / z_0$. For example: a total uplift 4 times smaller than the reference model (i.e., $U \cdot t_u / z_0 = 0.25$) will result in no lake if the reduction is performed on U, whereas if t_u is decreased, then a lake develops through nearly 1 m.y. Figure 7C shows the same results as in Figure 7A, but uses double the uplift duration than that of the reference model. Although the lake formation and closure remain invariant (relative to Fig. 7A), the adimensional times of opening and extinction of the lake are significantly delayed. However, it is interesting to remark that

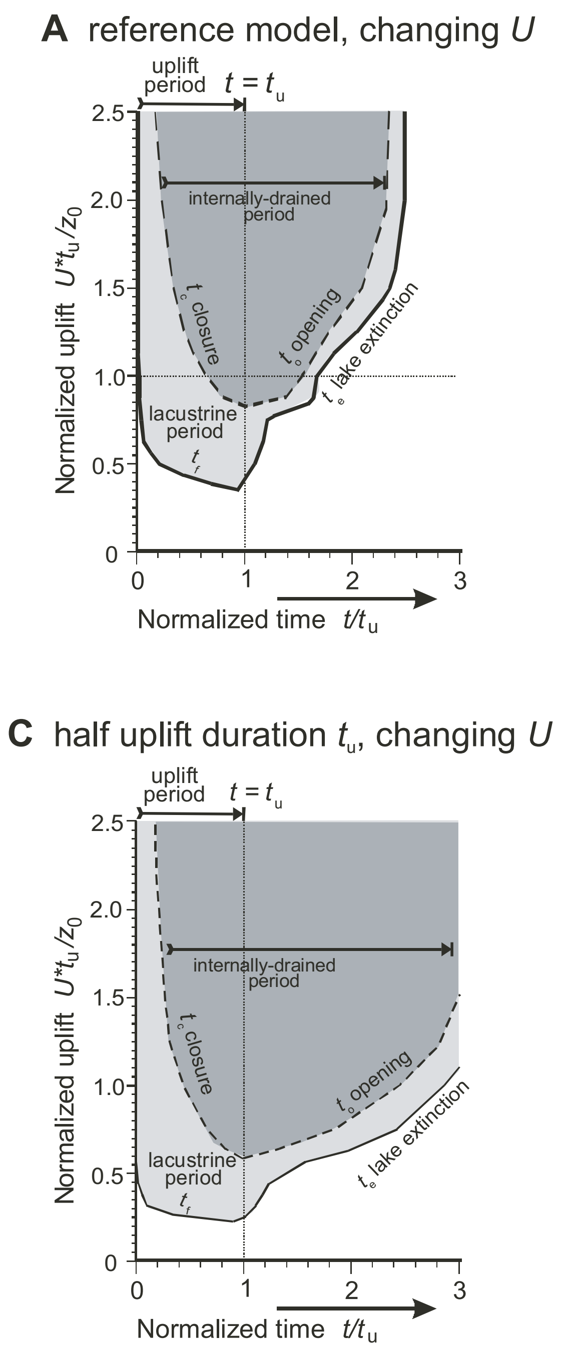


Figure 7. Effect of U, t_u, z_0 (uplift rate, uplift duration, and initial maximum topography) on lake evolution. The light shade indicates the existence of a lake; the dark shade indicates the lake is endorheic (closed drainage). The graphs show the lake evolution in adimensionalized time as a function of an adimensional total uplift. (A) Results of the reference model when changing the uplift rate; (B) same as A, but changing the duration of the uplift; (C) same as A for half uplift duration (0.5 m.y.). The differences between the plots indicate that it is not possible to find a precise, simple parameterization of the lake evolution as a function of these parameters. The relative similarity between them shows that, keeping the climatic and lithological parameters fixed, and only as a first approach, lake evolution is mostly controlled by the ratio between the uplift-generated topography $U \cdot t_u$ and the initial topography z_0. $U \cdot t_u / z_0 = 1$ (corresponding to the reference model) is plotted for reference.

the numerical experiments predict that, generally, an open lake survives for only a short time after uplift ceases, whereas an endorheic basin frequently lasts more than twice the duration of the uplift (see also Fig. 4).

Finally, the effects on this drainage evolution of the vertical motions related to lithospheric flexure were investigated, which were not incorporated in the models above. For this, the load/unload at every column related to sedimentation, erosion, and/or changes in water column were considered, and the lithospheric vertical motions $U_l(x)$ (equation 4) related to this surface mass redistribution were calculated. The reference model (Table 1) was then recalculated for different values of lithospheric elastic thickness (T_e). Large T_e values imply longer-wavelength and smaller-amplitude vertical motions. The results show (Fig. 6B) that post-tectonic uplift at the outlet, related to erosional rebound of the escarpment, induces a relevant delay on t_{hc} and t_e only for large lakes and low T_e. For a lake length of $l_l = 50$ km, relevant delay (larger than 10%) of t_{hc} and t_e (relative to $T_e = \infty$, i.e., no isostatic vertical movements) are predicted only for T_e values smaller than 7 and 11 km, respectively. In contrast, for $l_l = 200$ km, t_{hc} and t_e are significantly delayed for $T_e < 40$ km and $T_e < 100$ km, respectively (Fig. 3C). T_e values smaller than 3 km (for l_l = 50 km) and 20 km (for l_l = 200 km) produce an outlet rebound related to escarpment erosion that exceeds the erosion rate of the divide, thus delaying the opening of the basin during much larger periods of time.

INTERPRETATION AND DISCUSSION

The lake evolution model presented here assumes a 1D (along-stream) approach to water and sediment flow. In nature, water and sediment are collected in catchments with little symmetry across the river strike. Two-dimensional (planform) modeling was discarded for this study, not only for being extremely expensive in terms of computation time, but also to facilitate the reproducibility and interpretation of the results and because of the difficulty of designing a simple and universal setup (e.g., initial 3D topography). Besides, results from a planform model predict lake capture evolutions that are appreciably dependent on the resolution of the discretization grid (Garcia-Castellanos et al., 2003). A preliminary comparison with the results by Garcia-Castellanos et al. (2003) indicates that out-of-plane effects accelerate the capture process by a factor of ~2–4 (though this depends strongly on the initial 3D topography adopted for the model), maintaining the same qualitative effects of each parameter. Therefore, although the models shown here do provide a reliable quantification of the relative importance of each of the processes involved, the absolute values of lake timing obtained here must be viewed with caution.

A second and more fundamental limitation that must be kept in mind comes from the power-law model adopted for river incision and transport. The ongoing research on the appropriate equations describing these processes (e.g., Whipple and Tucker, 1999; van der Beek and Braun, 1999; Sklar and Dietrich, 2001) has not yet provided an approach with true prediction capability. The different forms proposed for the stream power law are often more sophisticated than that used here, incorporating to equation 1 a dependence on sediment load, granulometry, channel width, or climatic episodicity, for example. The laws proposed remain sometimes grounded on empirical relationships rather than on the physics of the involved processes, or are subjected to poorly constrained parameters such as lithology, vegetation, climatic episodicity, etc. Sets of m and n values (equation 1) different from those used here but still agreeing with field measurements would predict different velocities of upstream migration of erosion (e.g., Whipple and Tucker, 1999) and therefore lake capture. For the purposes of this paper, the simplified form adopted is sufficient, but the results presented here exemplify the extraordinary potential of the stream power law to improve our understanding of landscape evolution, and encourages the search for improved process-based models with increased predictability. Conversely, they suggest that a broad database giving constraints on the evolution of a large number of lakes may help us to refine the available surface process models using the techniques described in this paper.

According to the modeling results, the evolution of lakes is sensitive to the initial geometrical configuration, lithology, tectonic uplift, and climate in a similar degree, and all these factors can change the timing of lake evolution by several orders of magnitude. The most remarkable result is that evaporation, by reducing outlet water discharge in lakes, can substantially reduce erosion rates and extend the lake's life. Evaporation is therefore a key factor triggering and controlling the formation of internally drained basins. A good example for the stage of drainage closure is Lake Issyk-Kul (northern Tien-Shan, Kyrgyzstan), an ~180 × 50 km closed lake trapped at 1600 m above sea level in the forefront deformation of the Himalayan collision. Lake Issyk-Kul, with a P/E coefficient of 0.28 remains internally drained since the late Pleistocene (De Batist et al., 2002), although short periods of historical reopening have been reported. Together with the intense seismic activity in the region and the shortening velocities between the Indian and Asian domains, this suggests that, in geological time scales, Lake Issyk-Kul is presently being incorporated into the internally drained central Asia.

The topography upstream from the tectonic barrier is also a key factor controlling the reduction of water output through the lake's outlet, and it cannot be neglected in simulating the development of internally drained basins. All other parameters being constant, a relatively flat upstream topography implies that the lake attains a large area with small amounts of uplift, favoring evaporation and drainage closure, whereas a steep topography implies smaller lakes and less water evaporation for the same amount of uplift. This means that the formation of internally drained basins requires slower uplift rates in flat regions (e.g., Lake Bungunnia, S. Australia) than in mountainous regions (e.g., Lake Issyk-Kul, Himalayan collision zone).

The model successfully reproduces the notion that, after tectonic forcing ceases, lakes open and extinguish by a combination of sediment fill and outlet erosion (Fig. 2A). Lake elevation is the most relevant geometrical parameter controlling the timing of this process and dominates over the lake's length and depth (Fig. 6A). Thus, decreasing simultaneously by one order of magnitude all three geometrical parameters, capture slows down by a similar amount as a result of lower lake altitude and potential energy. In turn, if only length and depth are reduced and lake elevation is kept unchanged, capture is accelerated.

The flexural response of the lithosphere to surface mass redistribution (isostatic rebound and subsidence) also induces a scale-dependent effect. For the range of elastic thickness most frequently reported in continental lithosphere ($5 < T_e < 20$ km; Watts, 1992), lakes larger than 50–200 km have a post-tectonic evolution sensitive to lithospheric flexure. An important implication is that surface transport in large, closed lacustrine basins interplays with isostasy and erosion in the surrounding areas to control the long-term landscape evolution, and that the water budget in internal basins might have a key effect on the topographic and drainage evolution of continental margins (in addition to other factors such as isostatic rebound or escarpment retreat; Tucker and Slingerland, 1994). In particular, the peneplanation of Gondwana prior to breakup facilitated, together with the flexural uplift of the new continental margins, the formation of large, closed internal basins in southern Africa (Summerfield, 1991), triggering a period of intracontinental deposition lasting for most of the Cenozoic. The increase in altitude of these basins by sediment infill, possibly in combination with a wetter climatic phase, might have eventually triggered their drainage opening in late Tertiary times, in a similar way to that proposed by Garcia-Castellanos et al. (2003) for the Ebro Basin (NE Iberia).

In the last decade, numerical modeling techniques have revealed that the spatial distribution of erosion and surface transport controls the internal structure and tectonic evolution of orogens (e.g., Beaumont et al., 1992; Avouac and Burov, 1996; Willett, 1999; Persson et al., 2004). These studies have not explicitly addressed the effects of internally drained, endorheic basins on surface transport during orogenesis, despite the fact that these occupy large areas of many mountain belts on Earth (e.g., Andean Altiplano, Tibetan Plateau, and Tarim Basin). The results obtained in this work indicate that the hydrologic balance in intramountain basins determines their duration, possible closure, and the volume of their sedimentary infill. Long-lasting lacustrine basins are promoted and prolonged by low initial topographic gradients (low z_0), large and durable uplift rates (high U, t_u) acting over large areas (large l_u), dry climate (low P/E ratio), hard lithology (low K_f, K_d), and in the case of lakes larger than ~50 km, low lithospheric rigidities.

Orographic effects on precipitation (not explicitly incorporated to the model) block the inflow of humid air into intramountain areas (e.g., Roe et al., 2003), lowering the precipitation/evaporation ratio and thus facilitating endorheism, as shown in Figure 5. Accounting for this would in fact prolong the endorheic periods obtained with the model above, since the growing tectonic barrier would result in drier climate upstream from the uplift region. A longer endorheic period would imply larger sediment accumulation in the closed basin and lower erosion of its flanks, which in turn would induce a propagation of shortening toward the external parts of the orogen (e.g., Persson et al., 2004), reinforcing the closure of the basin by localizing precipitation further away from it. The ongoing drainage closure of Lake Issyk-Kul mentioned above, for example, may be not just a consequence but also a cause of the further propagation of tectonic deformation into Asia. Similarly, it is a remarkable concurrence of drainage closure in the Ebro Basin (NE Iberia) at 35 Ma and the dramatic reduction of tectonic deformation along the Pyrenees at ca. 30 Ma (Garcia-Castellanos et al., 2003, and references therein). This suggests that the coupling between orographic precipitation and tectonic deformation might be sufficient to explain the formation of long-lasting intramountain basins, such as the Neogene Ebro Basin, the Lake Issyk-Kul Basin, or the Andean Altiplano, and an inherited tectonic structure, such as a crustal-scale weakening, is not required. From the perspective of this feedback effect, the formation of an endorheic intramountain basin during the early stages of orogenesis, favored by a dry climatic setting and a low-relief inherited topography, might be reinforced and persist in the later tectonic and drainage evolution of the orogen. Future numerical models coupling the dynamics of tectonic, surface, and climatic processes should allow further investigation of the relevance of such effects.

CONCLUSIONS

Within the approaches inherent to the model described above, the following conclusions can be made:

1. The defeat of a river by tectonic uplift and the development of an internally drained basin are dependent on geometrical constraints (initial relief, length of the river), lithological parameters (rock erodibility), tectonics (uplift rate, duration, and its spatial distribution), and climate (precipitation and evaporation rates). For example, quantitatively determining the climatic conditions and uplift rate under which a particular lake and/or internally drained basin was formed requires information on the paleotopography upstream from the tectonic barrier (a flatter upstream topography would require less uplift to close drainage and vice versa). This ultimate goal, however, remains limited by our capability to accurately predict river incision and transport.
2. River defeat (uplift rate $U_T >$ river incision rate e) results in the formation of a lake, which only develops into an internally drained basin if (a) the ratio P/E (precipitation/evaporation) is significantly lower than 1; and (b) uplift persists until the lake is large enough to evaporate all collected water. In the notation used above, the necessary condition to generate an internally drained basin out of a defeated river is $l_l > l_b \cdot P/E$.
3. The drainage closure of a lake induces an important prolongation of the lake's life beyond the end of tectonism by

inhibiting erosion along the drainage outlet. The duration of the lake is increased by a time comparable to the duration of the endorheic period. This leads to a prolongation of the lake duration by a factor generally larger than 2, relative to a lake formed under similar conditions but not attaining endorheism.

4. The post-tectonic extinction of lakes larger than 50–200 km is significantly delayed by flexural isostatic uplift occurring in response to erosion at the topographic barrier.

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Daniel Garcia-Castellanos

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