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Late Neogene Erosion of the Alps: A Climate Driver?

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Key Words

exhumation, thermochronometry, sedimentation, orogeny

Abstract

As with many mountain belts globally, the Alps have seen a large increase in sediment yield in the late Neogene. The hypothesis that this increase results from climate-change impacts on erosion rate over the past ∼5 Ma is testable, given the extensive work completed in the Alps. Sediment budgets, thermochronology-based cooling rates, and estimates of modern rock uplift and erosion in the Western Alps are reviewed to search for correlation in space and time that might fingerprint climatic events. Major increases in sediment yield are apparent starting in the late Messinian (5.5 Mya) and accelerating toward the present day; this suggests the occurrence of a series of climatic changes that increased erosion rates, even as tectonic processes appear to have slowed or stopped. At present, tight correlations in time are difficult to establish and there remain open questions about the role of tectonics in the coupled system.

INTRODUCTION

The Alps have long served as a field laboratory for the testing of geological ideas from ocean formation to orogenesis, so it not surprising that the Alps have played an important role in the recent problem of assessing the impact of late Cenozoic climate change on mountains and mountain-belt erosion. The relationships between tectonics, erosion, and climate are complex with multiple feedbacks, but it is now widely accepted that increased climate variability and alpine glaciation associated with late Cenozoic global cooling has increased erosion rates in most of the world's orogenic belts (Donnelly 1982, Hay et al. 2002, Molnar 2004, Molnar & England 1990, Zhang et al. 2001). This is evident from the increase in sediment accumulation in oceanic and continental basins over the past ~5 Ma, even after accounting for the bias of ocean-floor subduction and recycling of sediment on continental margins by sea-level fluctuations (Hay et al. 1988). Although there is a cooling trend through most of the Cenozoic, since the mid-Pliocene the climate has become both colder and more variable with large-amplitude temperature fluctuations reflecting both eccentricity and precession forcing (Zachos et al. 2001). This cooling led to the onset of Northern Hemisphere glaciation ~2.5 Mya (Raymo 1994). Although it is difficult to time the onset of glaciation in a specific mountain range, presumably alpine glaciation was established or enhanced at this time in many mountain belts, and effective glacial erosion provides an attractive mechanism for the increased sediment flux from continental areas.

As a mid-latitude mountain range, the Alps have been strongly affected by Pleistocene glaciation. The modern morphology of the Alps is dominated by glacial features, including large, overdeepened valleys, many of which contain large postglacial lakes (**Figure 1**). In addition, there is both a historical and geological record of glacial advance and retreat over the Holocene and Pleistocene (2 Mya). Over a longer time span, the sediment yield from the Alps shows a large increase over the past ~5 Ma, which is consistent with global observations (Hay et al. 1992,

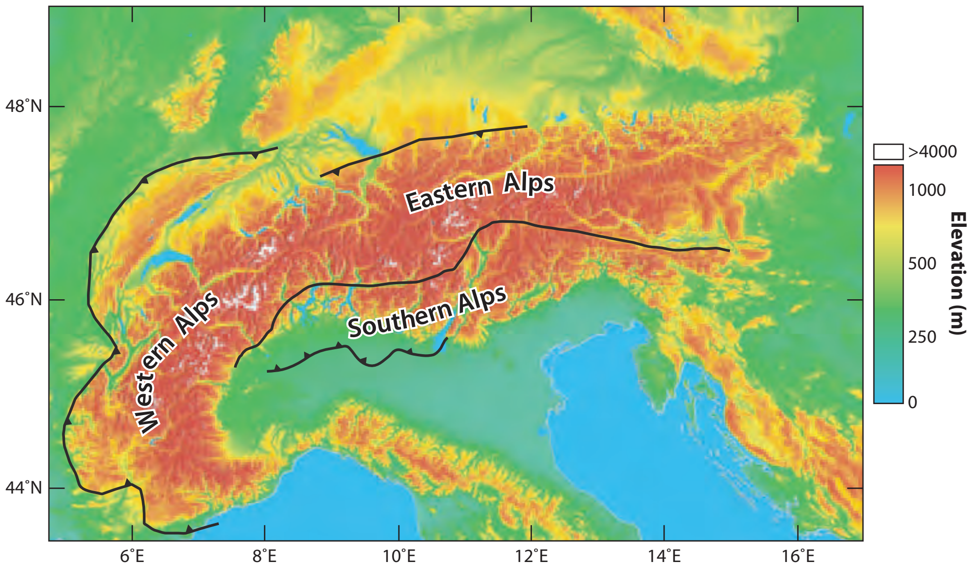

Figure 1

Topography of the European Alps with Western, Eastern, and Southern regions. The northern boundary of the Southern Alps corresponds to the Insubric and Periadriatic lines. The deformation fronts to Tertiary Alpine deformation are marked (*black lines*).

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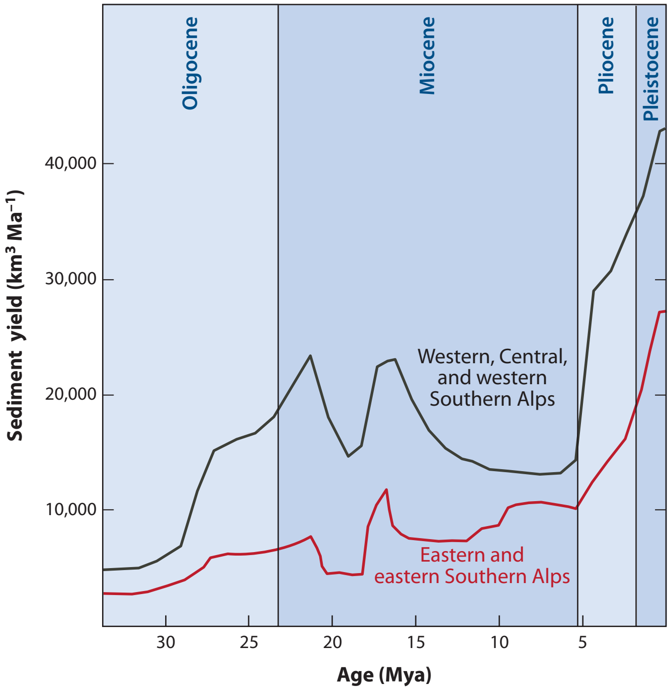

Figure 2

Rate of sediment production from the Alps with data from Kuhlemann (2000). Source areas are defined to separate the Southern Alps into eastern and western regions with a boundary at approximately 10°E. This differs from Kuhlemann et al. (2002), whose authors included the entire Southern Alps in the Western Alps yield curve. Note the increase in yield starting ~5 Mya and increasing to the present day.

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Kuhlemann et al. 2001b) (**Figure 2**). Given their accelerated sediment yield and the clear impact of glaciation, the Alps have served as an example of how the intensification of climate variability and glaciation affects an orogenic belt (Champagnac et al. 2007, Molnar 2004).

However, the resolution provided by the detailed studies of sediment yield from the Alps, particularly the detailed work of Kuhlemann (Kuhlemann et al. 2001b,c; 2002), demonstrates that the yield has variations in both space and time. Most important, the increase in sediment yield occurred too early to correspond exclusively to Pleistocene glaciation. In the Southern Alps and Po Foreland Basin, where age control is best, it appears to have begun in the latest Miocene (5.5 Mya) (Willett et al. 2006), but other basins around the Alps also show an increase ~5 Mya. The entire Plio-Pleistocene is a period of rapid climatic change, at least globally, so this increase in sediment flux could still be climatically driven; but if so, it is not exclusively a signal of glacial erosion.

In any case, the correlation between the sediment-yield curve (**Figure 2**) and global temperature proxies has stimulated much recent interest and research. The longer timescale over which sediment production has accelerated opens other opportunities to measure the processes of enhanced erosion. In particular, low-temperature thermochronometry, which provides a measure of near-surface cooling rates, should be capable of resolving accelerations in exhumation over the million-year timescale implied by **Figure 2**. A number of local and regional studies have targeted late Neogene exhumation rates (Vernon et al. 2008). Although generally confirming late rapid exhumation, some surprising spatial and temporal variations have been identified; these are discussed in this paper.

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One complication with the Alps, as with all active orogens, is differentiating erosion-rate changes driven by climate change from those associated with tectonic activity. The late Neogene tectonic state of the Alps is difficult to establish because of the diminishing rates of convergence and deformation. It is clear that modern tectonic convergence in the Alps is a small fraction of its Tertiary average (Calais et al. 2002, D'Agostino et al. 2008). In addition, the pole of rotation for the relative motion between Adria and Europe has shifted toward the east, so the Western Alps are no longer undergoing any net contraction (Delacou et al. 2004). The Eastern Alps continue to shorten, albeit at a slow rate with a relative plate motion of less than 2 mm year^{-1} (D'Agostino et al. 2008). However, these rates cannot be extrapolated back over the 5-Ma time period of interest. Shortening was clearly occurring on both the northern and southern margins of the central Alps in the late Miocene, and possibly into the Pliocene or even later (Fantoni et al. 2001, Ustaszewski & Schmid 2007). Deformation in the Alpine interior is more difficult to resolve independently of the erosion rates that we are trying to determine, but there is at least circumstantial evidence of Pliocene-to-present deformation from high—and differential—exhumation rates measured in the internal and external massifs (Fugenschuh & Schmid 2003, Malusa & Vezzoli 2006, Persaud & Pfiffner 2004).

The tectonic activity is important as a potential explanation of both temporal and spatial variability in exhumation and sediment supply, but also because it affects the way in which a mountain belt responds to changing climate and erosional conditions. A tectonically inactive mountain belt responds to accelerated erosion by more rapid isostatic uplift, which might enhance local relief but does not suffice to restore elevation to pre-erosion levels (Molnar & England 1990). On the other hand, an active mountain belt can respond to enhanced erosion by a shift in deformation so that topography is restored or maintained, but with higher exhumation rates over a more limited spatial area (Stolar et al. 2006, 2007; Whipple & Meade 2004, 2006). These different responses predict different sediment fluxes and can serve to distinguish between active and inactive orogens. The Alps could prove to be a test case for such models.

In this paper, I review the case for accelerated erosion of the Alps over the past 5 Ma. I start with a review of sediment-yield studies that have motivated much of this interest. Then I examine theoretical considerations of the orogenic response to accelerated erosion, which provide a useful framework in which to consider observations. I then present evidence for accelerated erosion from other sources, including tectonics, sedimentology, thermochronometry, and cosmogenic isotope-based estimates of erosion rates. The focus is on the past 10 Ma of Alpine history, when climatic change has been the most dramatic. Although some reference to Eastern Alps data is made, the focus is necessarily on the Western Alps in Switzerland, France, and northern Italy, as this is where most of the relevant work has been done to date. Finally, I discuss the commonalities and differences in these data, correlations, and the relationship to paleoclimate data.

NEOGENE SEDIMENT YIELD FROM THE ALPS

Like most mid-latitude mountain belts (Molnar 2004, Zhang et al. 2001), the Alps exhibit an increase in sediment yield over the past 5 Ma. However, unlike most mountain belts, the Alps have seen sufficient work on the sediment-yield record to resolve this increase in time and in space, primarily through the work of Kuhlemann (Kuhlemann 2000, 2007; Kuhlemann et al. 2001a,b; 2002). Large uncertainty remains with such data. The time resolution is limited by stratigraphic age control in the depositional basins. The spatial resolution is limited by the need to reconstruct past source-area drainage basins (Kuhlemann et al. 2001c, Schlunegger & Hinderer 2001). However, the primary error comes from trying to estimate sediment volumes that have been tectonically accreted into orogens or subducted—in the Apennines, for example, where

Alpine-sourced basins comprise much of the orogen, including at depth where their geometry cannot be assessed. Nonetheless, there is a clear signal of increased sediment yield from the Alps over the past 5 Ma (**Figure 2**), which suggests an apparent acceleration of erosion rates toward the present day.

MSC: Messinian salinity crisis

Some investigators suggest that the timing of this accelerated erosion rate varies across the Alps. In particular, Kuhlemann et al. (2001b) differentiated the onset of accelerated erosion in the Western and Eastern Alps, finding that the increase started 5 Mya in the Western Alps but 3.5 Mya in the Eastern Alps. This is an important observation because climate change–driven erosion is unlikely to be highly spatially variable over a small mountain range like the Alps. However, this difference in timing is difficult to resolve in the source areas. The major drainages to the north (Rhine), southwest (Rhone), and south (Po-Adriatic Basin) show an increase in sediment yield ~5 Mya, whereas the Danube system seems not to increase until 3.5 Mya. The Eastern Alps, defined geologically to reside north of the Periadriatic Line, are almost entirely in the Danube drainage basin (**Figure 1**) and thus seem to have experienced the increase in erosion rate later than the Western Alps. However, the eastern Southern Alps, which reside at the same longitude but south of the Periadriatic Line, shed sediment to the Adriatic Basin, so their erosional history cannot be separated from that of the Western Alps. For this reason, Kuhlemann et al. (2001b) included the entire Southern Alps with the Western Alps sediment yield. **Figure 2** shows the data of Kuhlemann et al. (2001b), but with the eastern Southern Alps included with the rest of the Eastern Alps. With this presentation, the difference in time of increase of erosion rate from west to east largely disappears, and the entire Alps appears to have experienced accelerated erosion rates approximately 5 Mya.

However, this illustrates more the resolution problems of sediment-yield data rather than any robust signal of increased sediment yield because the result depends entirely on how one partitions sediment in the Po-Adriatic Basin to source areas in the Western Alps, the Eastern Alps, the Apennines, and the Dinarides (Bartolini et al. 1996, Kuhlemann 2000, Mancin et al. 2009, Willett et al. 2006). There is little basis for establishing provenance of Adriatic sediments of this age, so the data above are partitioned primarily by contributing area and by analogy with modern sediment supply. Partitioning source areas in the Alps and Apennines is particularly difficult because the Apennines are composed primarily of Miocene clastics shed from the Alps and so have the same original provenance.

Temporal control in the Adriatic is good, with a well-defined chronostratigraphy, so the onset of rapid erosion is well timed. In fact, the onset of high erosion rates can be extremely well resolved, as it appears to have occurred in the late Miocene during the Messinian Salinity Crisis (MSC), which has been well studied and dated (Krijgsman et al. 1999). Toward the end of the MSC, basins around the Mediterranean, including the Adriatic, received a large influx of coarse, siliciclastic sediment deposited in fluvial-deltaic environments or brackish-water lagoonal environments. The deposition of this so-called Lago Mare facies sediment clearly postdates the deposition of the primary MSC evaporites, as well as the intra-MSC unconformity that marks the dramatic lowering of Mediterranean sea level ~5.5 Mya. In addition, it predates the reestablishment of marine conditions at the Miocene-Pliocene boundary 5.33 Mya (Roveri et al. 2008). It thus records an important Mediterranean-wide depositional event. The volume of siliciclastic sediment in the Po Basin associated with this event is quite large (Fantoni et al. 2001, Roveri et al. 1998), particularly given the depositional interval of only 0.17 Ma. It reflects nearly an order-of-magnitude increase in sediment yield (Willett et al. 2006). This increase likely results from a combination of factors, including the disequilibrium of the landscape owing to the kilometer-scale, base-level fall of the Mediterranean and the creation of accommodation space on the Mediterranean margins as the water level rose. However, this increase in local sea level does not arise from marine incursion

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because the facies are clearly brackish water. Therefore, the increase must represent an additional fresh water input, which implies a climatic change including higher average precipitation rates—an inference also supported by the high-energy characteristics of the siliciclastic depositional environment (Keogh & Butler 1999, Rouchy et al. 2001, Roveri et al. 2008, Willett et al. 2006).

At the Miocene-Pliocene boundary, marine conditions were restored to the Mediterranean. However, sediment yield remained high through the Pliocene in the Adriatic, Rhone, and Rhine drainage systems. Approximately 3.5 Mya, sediment yield again increased throughout the entire Alpine region and continued to accelerate through the Pleistocene, although temporal resolution is not sufficient to note abrupt changes in rate.

ISOSTATIC RESPONSE TO ENHANCED EROSION

The tectonic response to enhanced mountain erosion includes both vertical and horizontal motions as well as changes in the distribution of both deformation and exhumation rates. However, these responses vary depending on the tectonic setting, and in convergent orogens, as we consider here, it depends particularly on whether a mountain belt is active. I consider these two cases separately.

Passive Response

If an orogen is no longer tectonically active, the response to enhanced erosion is primarily a vertical, isostatic one. Removal of material by erosion leads to isostatic uplift of the underlying lithospheric column. Given typical crust and mantle density, the responding uplift is approximately 80% of the eroded crustal thickness (Molnar & England 1990). This proportionality also holds for rate calculations, so that an increase in surface-erosion rate should be matched by a corresponding 80% increase in rock-uplift rate. Climate-driven changes in erosion rate should thus map directly into total sediment-production rate as well as exhumation rates. Rock-uplift rate is always a fraction of the erosion rate, so topography is slowly destroyed by erosion. In addition, under constant climatic conditions, erosion rates and sediment-production rates will decrease with time over a timescale controlled by the erosion rates and the excess thickness of the crust.

An isostatic response to erosion can be associated with an important change of relief. If a low-relief topography is incised by deep valleys, the resultant isostatic rebound will elevate the neighboring peaks—a principle that has been long recognized as important in mountain belts (Wager 1933). Molnar & England (1990) argued that late Cenozoic glaciation of alpine mountain belts would produce such peak uplift if glacial erosion were effective at carving and flattening deep valleys down to near sea level (**Figure 3*a***). If glacial erosion removes an amount of rock equivalent to a uniform layer of thickness h, the resulting topography would have a mean elevation of $0.8h$ but peak elevations of approximately $1.8h$.

In the Alps, this idea was pursued by Champagnac et al. (2007). Assuming that Alpine valleys were completely eroded at some time in the recent past, they calculated the isostatic uplift owing to the mass removed from the valleys. Adding a background erosion rate and assuming most material was removed over the past 1 Ma produces a pattern of uplift with the maximum in the center of the Alps, similar to modern geodetic-based uplift rates (Gubler et al. 1981). However, the maximum isostatic uplift rate is $\sim$0.6 mm year^{-1}, which is approximately only half the magnitude of the present geodetic-based uplift rate. This suggests that an increase in relief by valley carving could be a significant component of modern uplift rates but is insufficient to fully account for them.

There is also a regional response to enhanced erosion if topography is regionally supported by an elastic lithosphere. In this case, destruction of topography produces long-wavelength rebound that is most apparent on the margins of an orogen where the downward flexing of the lithosphere

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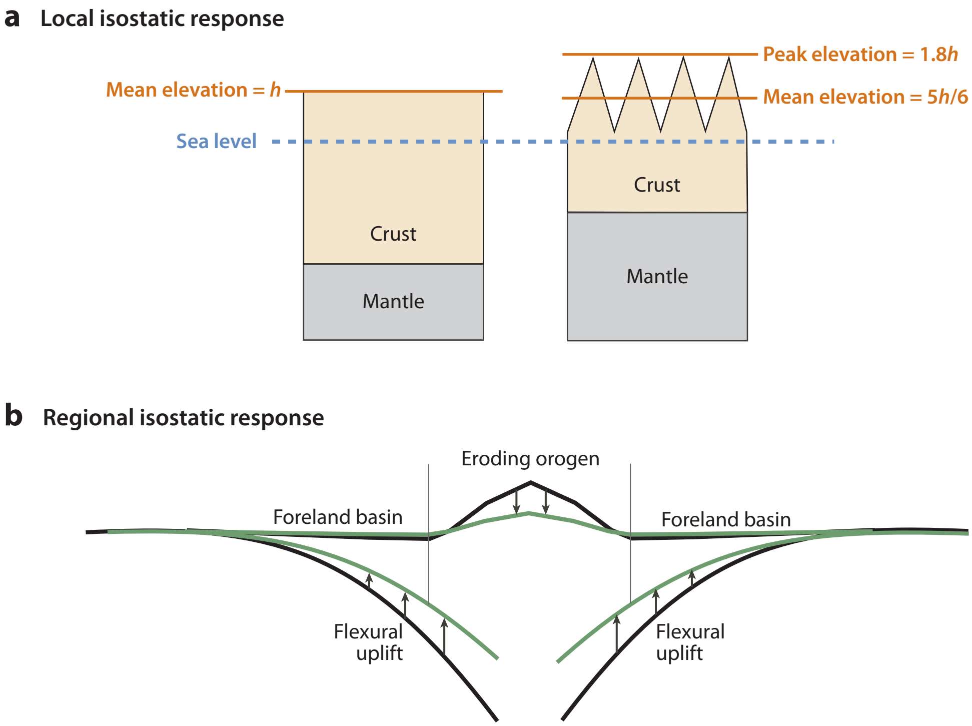

Figure 3

Isostatic response to enhanced erosion. (*a*) Local isostatic response to erosion of h and an initial elevation of h. Crust rebounds by $5h/6$ so that topographic remnants can reach elevations of up to $1.8h$. (*b*) Regional response where orogenic load is flexurally compensated, forming foreland basins. Decrease of load leads to uplift and erosion of the neighboring basins.

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produces a foredeep or foreland basin (**Figure 3*b***). A net reduction in orogenic topography unloads the lithospheric plate and produces rebound of the adjacent foreland basin.

This process is inferred for the Alps based on the depositional and erosional history of the Alpine foreland basins. The Northern Alpine Foreland Basin (NAFB) has a well-documented history of subsidence and filling from early, underfilled conditions in the Oligocene through most of the Miocene (Homewood et al. 1986, Pfiffner 1986, Sinclair & Allen 1992). However, the youngest sediments in the basin vary from approximately 11 Ma in western Switzerland (Kälin 1997) to as young as 8 Ma in Austria (Kuhlemann & Kempf 2002), indicating uplift and erosion of the basin after the Miocene. Subsidence analysis demonstrates this reversal from subsidence to uplift (Genser et al. 2007), and studies of thermal indicators such as vitrinite reflectance provide estimates of 2 to 4 km for the total erosion (Mazurek et al. 2006, Schegg 1993). Age control for onset of this erosion comes from fission-track data from exploration wells in the basin. In a study of apatite from three wells in the Swiss region of the NAFB, Cederbom et al. (2004) estimated the onset of cooling between 5.5 and 4.8 Mya. Other wells to the east and west of this initial study yield similar onset times (Cederbom et al. 2008).

The southern foreland to the central Alps is the Po Basin, which has continued to accumulate sediment to the present day. Although this suggests that the orogenic load continues to increase, the depositional environment of sediments in the Po Basin indicate a shallowing upward trend with a transition from marine to terrestrial deposition in the early to middle Pleistocene. This

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indicates progressive filling of the basin, arising in part from the closure of the basin between the Alps and the Apennines (Bertotti et al. 1998). Interestingly, after corrected for the loading of the overlying sediments, the youngest marine sediments in the subsurface are presently above sea level (Scardia et al. 2006). This indicates that in spite of the continuing sedimentation, the Po Basin basement has been experiencing uplift relative to sea level. Scardia et al. (2006) place this uplift in the middle to late Pleistocene; the earlier uplift or subsidence history is difficult to resolve because the sediments are exclusively marine and water-depth information is insufficiently precise.

In the westernmost Alps, in the foreland basin of France, there is also evidence of flexural unloading. Champagnac et al. (2008) demonstrated tilting of up to 0.8% of foreland surfaces that are 1.8 Ma old, indicating foreland rebound that results from unloading since the Pliocene.

Active Tectonic Response to Enhanced Erosion

In addition to the vertical motions associated with isostatic rebound, an active orogen may respond to enhanced erosion by a redistribution of deformation. In particular, orogens that have attained a characteristic topographic form will deform to maintain that form independent of surface modification by erosion. The classic example of this behavior is the response of a critical orogenic wedge to erosion (Barr & Dahlen 1989, Barr et al. 1991, Dahlen & Barr 1989, Roe et al. 2006, Stolar et al. 2007, Whipple 2009, Whipple & Meade 2004, Willett 1999). An orogenic wedge deforming by a plastic deformational mechanism with a Coulomb yield stress will attain a constant surface slope, provided that the internal yield strength of the wedge and its basal detachment are constant and that convergence keeps the wedge everywhere at failure. If there is surface erosion, internal deformation will compensate to maintain the critical surface slope regardless of the distribution of surface erosion. The erosion rate distribution, in turn, depends on rock erodibility, precipitation distribution, and the river network's geometry (Stolar et al. 2007). The consequence is an orogenic system that tends toward a steady state in which erosion rates are matched by rock-uplift rates and the mean average surface of the orogenic wedge has a constant slope in the convergence direction (Willett & Brandon 2002). With variations in rock rheology or yield strength, the upper surface may not have a constant mean slope, but the principle of a critical form holds as long as gravitational stresses are an important component of the force balance.

Under these conditions, an important balance is maintained between the mass flux into an orogen resulting from convergence and the erosional flux from the orogen (**Figure 4*a***). These two fluxes tend to balance; the controlling characteristic of the orogen is the overall size, as characterized by the orogen width in the convergence direction W (**Figure 4**). Erosion rates and fluxes increase with the size and relief of an orogen, so change in orogen size provides the feedback mechanism that damps the system toward a steady state. If the erosional flux exceeds the accretionary flux, the orogen enters a destructive state in which the size decreases, thereby decreasing the erosional flux until the flux balance is again achieved (Jamieson & Beaumont 1988, Stolar et al. 2006, Whipple & Meade 2006). Conversely, a constructive orogen is one in which the accretionary flux exceeds the erosional flux and the orogen increases in size.

There are a number of characteristics of an orogen not in steady state. For example, an orogen in a destructive state as a consequence of an increase in precipitation rate will experience a decrease in the actively deforming width (**Figure 4**). This implies abandonment of active structures at the retro-deformation front (shown on the right in **Figure 4*a***) as deformation retreats into the orogen interior. In addition, there is an initial increase in sediment flux and average rock-uplift rate (**Figure 4*b***). However, as the width of the orogen decreases with time, the height and relief correspondingly decrease, leading to a decrease in erosional flux. Therefore, the new steady state is achieved with precisely the same flux as in the initial condition. The mean rock-uplift rate and

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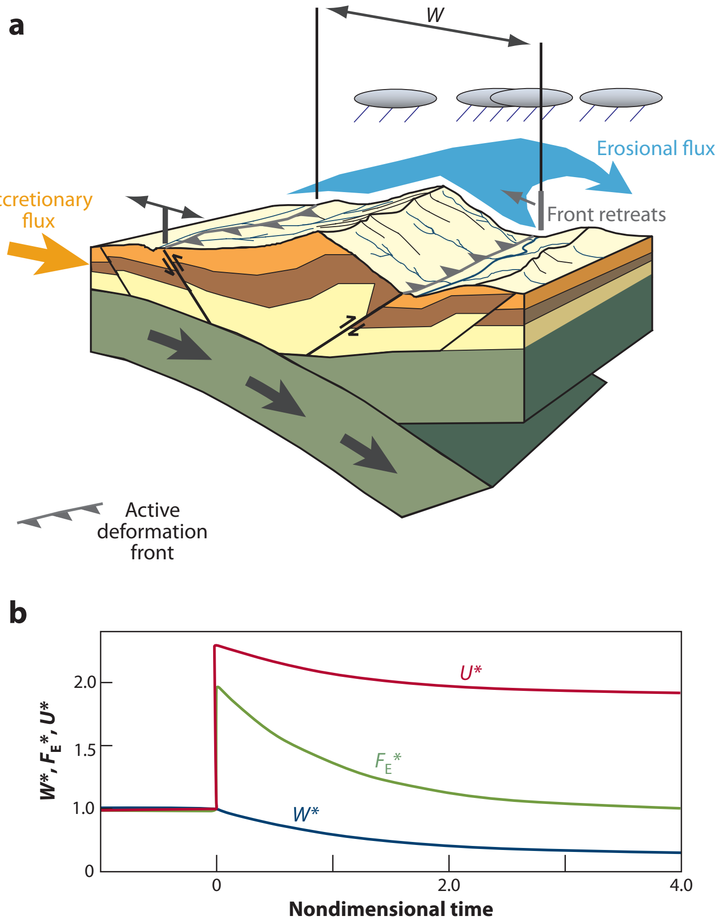

Figure 4

(*a*) Orogenic wedge model with accretionary flux (*orange arrow*) driving growth of the wedge, countered by erosional flux (*blue arrow*). When erosional flux exceeds accretionary flux, wedge width W decreases. Modified from Willett et al. (2006). (*b*) Response of wedge characteristics to a step increase in precipitation rate. Wedge width W^* decreases monotonically, but erosional flux $F_E{}^*$ and average rock-uplift rate U^* show an initial increase followed by a subsequent decline. All quantities are normalized by their initial values. Time is normalized by the characteristic time of the system. Modified from Whipple (2009).

the exhumation rate at steady state are higher because this flux is driven through a smaller area of Earth's surface, reflecting the smaller W. The new steady state is thus characterized by no net change in sediment yield, but a higher exhumation rate.

Willett et al. (2006) postulated this behavior for the Western Alps associated with the observed 5-Ma increase in sediment yield. The principal evidence for this is the abandonment of the southernmost Alpine-age folds and faults of the Lombardic thrust belt in the Po Plain. This event is well-timed to the end of the Messinian (5.3 Mya) (Fantoni et al. 2001, Pieri & Groppi 1981, Roure et al. 1990). The coincidence of the end of deformation in the Lombardic belt with the increase in sediment yield is a key observation in the inference of a climate driver to this event. There is secondary evidence from the north side of the Alps. The thin-skinned deformation in the Jura mountains, which marks the northernmost Alpine deformation front in the central Alps, also stopped around the end of the Miocene, although timing of this event is less precise (Becker 2000, Hindle 2008, Kälin 1997). However, if the uplift of the foreland basin, as discussed above, is mechanistically related to the cessation of shortening in the Jura (Willett & Schlunegger 2009), the timing on basin inversion is also a constraint on abandonment of this deformation front, which therefore must have occurred approximately 5 Mya (Cederbom et al. 2004).

The timescale of the transient response to enhanced erosion can be estimated based on critical wedge theory. The characteristic time—as used to scale **Figure 4**, for example—depends primarily on the erosional efficiency of the final state and only weakly on the size of the orogen and the uplift rate (Whipple & Meade 2006). It is difficult to directly estimate the erosional efficiency for a specific orogen such as the Alps, but we can estimate the response time by considering the perturbation to the system. Let us assume that the Swiss Alps were in a steady state in the late Miocene with a width of 220 km, a cross-sectional area of 3600 km^2 (Schmid et al. 2004), a convergence rate of 5 mm $year^{-1}$, and an accreted crustal thickness of 20 km. In that case, a doubling of the precipitation rate would produce a perturbation with a characteristic time of 24 to 30 Ma, where the range reflects variation in the initial precipitation rate. This suggests that any climatically driven change in erosion rate in the past 5 Ma has not equilibrated and that we would be in the early stages of the transient response with a nondimensional time of less than 1.0 (**Figure 4*b***). The corresponding steady-state wedge width is 160 to 180 km.

HOLOCENE EROSION RATES

Resolving rock-uplift rates and exhumation rates in space and time is an important component of distinguishing the type of tectonic response exhibited by an orogen and of establishing correlation to tectonic and climatic forcings. The modern system can be studied with a variety of techniques, although care of the relevant timescale of measurement must be taken. The Alps have seen significant geodetic leveling work over the past 100 years, and these studies give a good constraint on the rock-uplift rate at the decade timescale. In Switzerland, a clear pattern of uplift correlated to elevation has been established (Gubler et al. 1981, Kahle et al. 1997, Schlatter et al. 2005). Rock-uplift rates reach a maximum of more than 1 mm $year^{-1}$ in the central Swiss Alps but decrease toward the forelands (**Figure 5*c***). This is not very diagnostic in its own right, but the relationship to erosional unloading could be. There is one complication: The essentially instantaneous geodetic uplift rate will respond to the current erosion rates but also might contain a time-dependent component that reflects a viscoelastic response to past erosion rates or even ice loading from the last glaciation (Barletta et al. 2006, Stocchi et al. 2009).

Champagnac et al. (2007) argued that a significant fraction of the modern uplift arises from the recent carving of the deep Alpine valleys. However, given that erosion rates were already high—as early as the early Pliocene (5 to 3 Mya)—it is unlikely that all relief is a recent development, as they assumed. If the modern rock-uplift pattern is a passive isostatic response to erosional unloading, it is more likely a reflection of the distribution of erosion rate.

The Holocene erosion rates have been estimated from sediment loads in modern rivers and from lake-filling rates (Hinderer 2001). Both techniques give biased results because they can capture only the postglacial sediment production rates. Erosion and sediment production is cyclic over the glacial cycle, reflecting glacial and nonglacial conditions. Hinderer (2001) found a mean postglacial erosion rate of 0.6 mm $year^{-1}$, with the highest rates in the Rhone and Rhine drainage basins that cover much of the high Western Alps. Rates are lower in the Eastern Alps and peripheral regions. Modern erosion rates based on river sediment loads were considerably smaller, with a mean rate of 0.125 mm $year^{-1}$, illustrating a progressive decrease in sediment release as the postglacial cycle progresses.

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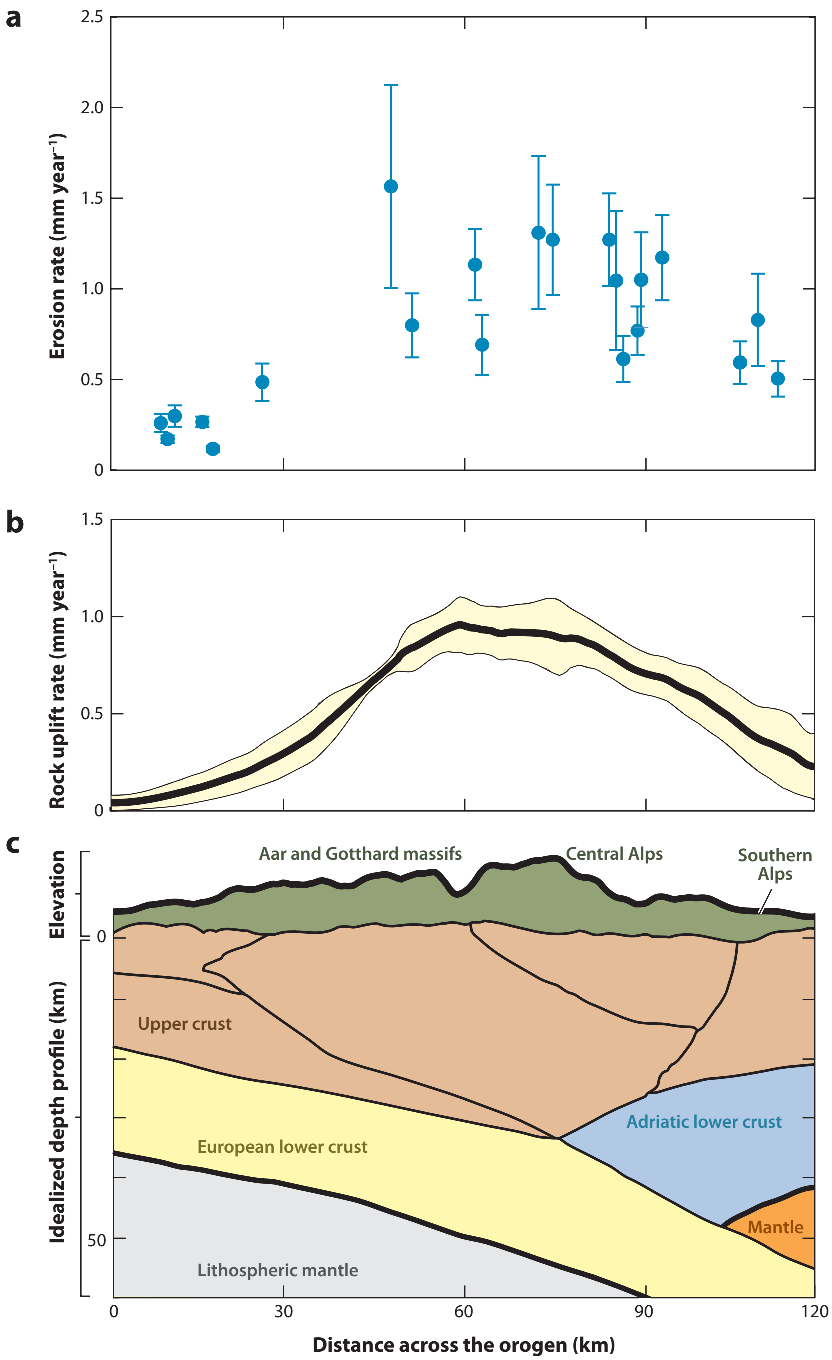

Figure 5

North–south transect across the Swiss Alps, showing: (*a*) cosmogenic ^{10}Be-based erosion rate averaged over the past few thousand years; (*b*) geodetic-based rock-uplift rate averaged over the past few decades; and (*c*) topography and crustal structure. Modified from Wittmann et al. (2007).

Higher resolution in space and time is provided by recent work that used cosmogenic isotope concentrations in modern sediments (Norton et al. 2008, Wittmann et al. 2007). By measuring ^{10}Be concentrations in modern sediments derived from drainage basins across the Alps, Wittmann et al. (2007) were able to construct an erosion map of the central Swiss Alps relevant on a timescale from 400 to 8000 years. Interestingly, these erosion rates match the geodetic rock-uplift rate in both pattern and magnitude (**Figure 5**). As discussed above, if rock uplift is purely an isostatic response to erosion, the pattern should be equivalent, but the ratio of rock-uplift rate to erosion rate should be approximately 0.8. Wittmann et al. (2007) found a ratio of 1.0. Based on the same data but a slightly different analysis, Champagnac et al. (2009) found a ratio of 0.9. The uncertainty on this calculation is more than 10%, so data obtained to date cannot distinguish between a passive isostatic response and a steady-state response in which erosion rates are equal to rock-uplift rates.

There seems to be little change in the erosion pattern from the long term (Ma timescale), which is discussed below, to the short term (Holocene). This makes it difficult to identify changing erosional or tectonic processes (Malusa et al. 2005a, Vernon et al. 2009a), but on the other hand does not require that these processes remain constant with time.

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THERMOCHRONOMETRY AND EXHUMATION RATES

Thermochronometric data provide an important constraint on long-term (Ma) erosion rates, and they potentially show changes in erosion rate given their ability to resolve both cooling rates and timing. Resolving changes in erosion rate over the past 5 Ma, given the moderate rates of erosion in the Alps, is best done through application of low-closure-temperature systems such as zircon fission-track dating, apatite fission-track dating, and (U-Th)/He dating of apatite and zircon. Many studies have used thermochronometry to constrain tectonic processes in the Alps, but it is only recently that studies have attempted explicitly to estimate post-Miocene erosion rates. Many of the early tectonic studies have produced data that pertain to late-stage erosion, and these data have been compiled and analyzed in various studies (Fugenschuh & Schmid 2003, Luth & Willingshofer 2008, Malusa et al. 2005b, Rahn 2001, Schlunegger & Willett 1999, Seward & Mancktelow 1994, Vernon et al. 2008, Wagner et al. 1977).

To establish changes in erosion rate, particularly as averaged over large regions, detrital thermochronometry is often a preferred method (Bernet et al. 2004b, Bernet & Garver 2005, Garver et al. 1999). A number of detrital studies have been conducted on Alpine sediment (e.g., Dunkl et al. 2001, Spiegel et al. 2000), but in particular, Bernet et al. made a major assessment of detrital zircon ages from a variety of late Tertiary sedimentary repositories (Bernet et al. 2001; 2004a,b; 2009). Interestingly, they were unable to resolve any differences in peak exhumation rates over the past 30 Ma of Alpine history. By isolating the youngest population of zircons in Alpine sediments and noting the lag age between the fission-track cooling age and the depositional age, they obtained the average erosion rate for the young population of zircons. Bernet et al. (2001) found that this lag age was consistently 8 Ma for the time interval of 30 Mya to the present day. This stands in stark contrast to the sediment-yield evidence for higher erosion rates in the past 5 Ma. However, it is important to note that the detrital zircon lag age does not reflect the spatially averaged erosion rate for the entire source area. Rather, the finding provides the average erosion rate for the fastest exhumed zircons. The implication is that the regions that experienced accelerated exhumation since the Miocene are not the same regions presently exhuming the youngest zircons.

Basement dating also has the potential to resolve changes in erosion rate, particularly through the use of multiple methods with different closure temperatures or through the construction of age-elevation relationships, which can determine an exhumation rate independent of the depth to the closure isotherm of the system. The most recent, and complete, analysis of basement fission-track data from the Western Alps was done by Vernon et al. (2008), who compiled apatite and zircon fission-track ages from more than 50 publications and additional unpublished studies. Their study focused on the Western Alps, west of ~9°E, because the deeper exhumation and lithology of the Western Alps makes them more conducive to thermochronometric study.

Regions that do have good spatial coverage, particularly with both apatite and zircon ages, can be analyzed for the average cooling rate, or more importantly, for changes in cooling rate derived from the difference in closure of the apatite and zircon systems. The zircon fission-track age and the apatite fission-track age give two points in time-temperature space, which, combined with the present-day surface temperature, can resolve an acceleration in cooling or erosion rate. **Figure 6** shows a summary of the acceleration in erosion rate derived from fission-track ages from

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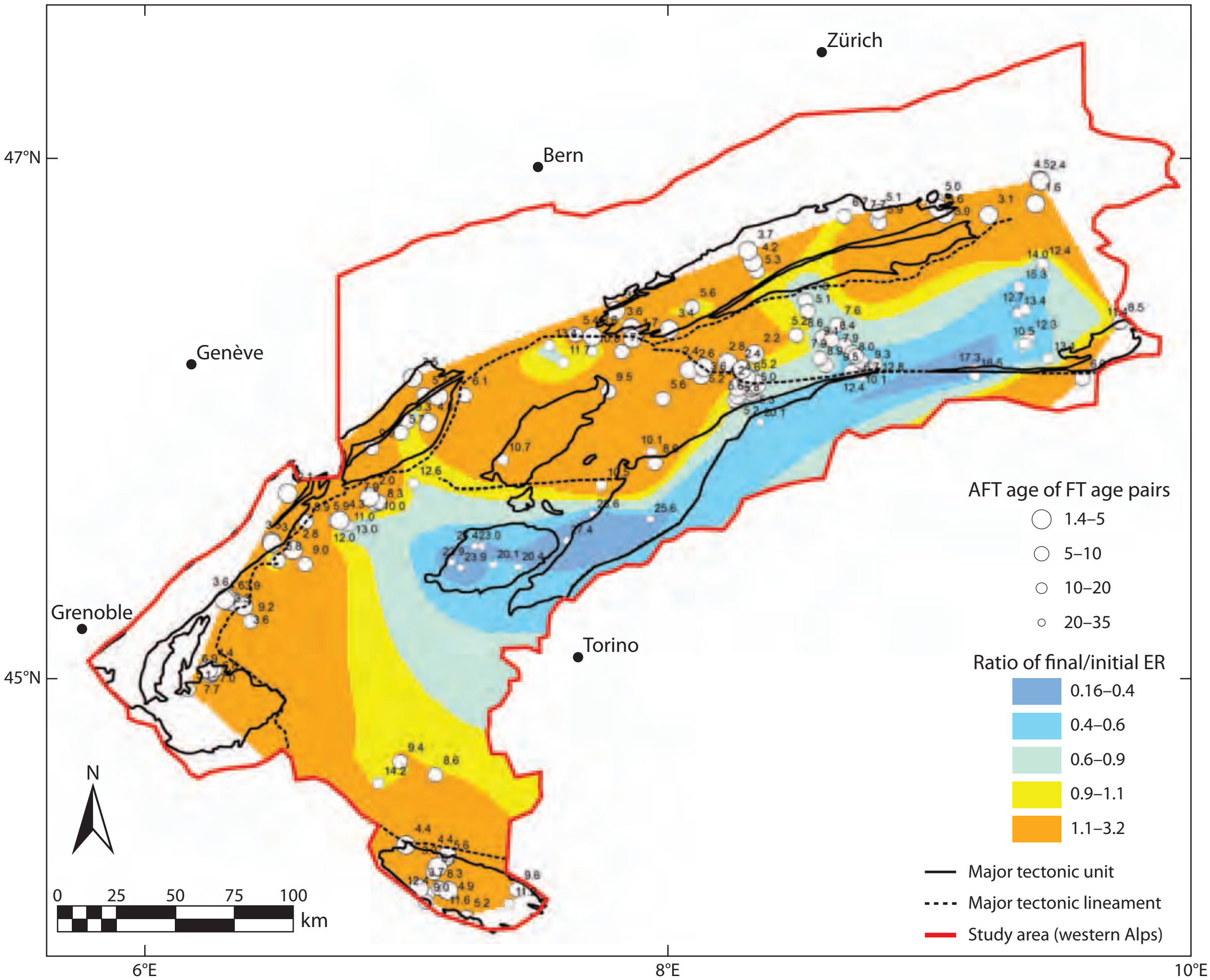

Figure 6

Ratio of two cooling rates. The first is derived from the time between closure of the zircon fission-track system and the apatite fission-track system; the second is from the closure of the apatite system to the present day. Values greater than 1 thus correspond to an acceleration in exhumation rate toward the present day. Abbreviations: AFT, apatite fission track; FT, fission track; ER, erosion rate. Taken from Vernon et al. 2008.

Vernon et al. (2008). To derive erosion rates, a one-dimensional thermal model was used to calculate closure temperature and the depth to the closure isotherm, including the effects of vertical heat advection. The erosion rate was then calculated from the measured age and the depth to the respective closure isotherm. The blue colors in **Figure 6** represent regions that have experienced a decrease in erosion rate; the orange-colored regions have experienced an increase in erosion rate. This is not necessarily a change at a single point in time. In fact, the transition is only resolved to the apatite fission track (**Figure 6**) (Vernon et al. 2008). However, given the range of apatite fission-track ages, most of the change occurred in the past 10 Ma. On average, most of the Western Alps experienced an acceleration in exhumation toward the present day. The exhumation rates over the past 4 Ma are roughly correlated with the modern rock-uplift rates, which also show the highest values in the north, external Alps (Vernon et al. 2008, 2009a). In addition, there is an interesting pattern of decreasing rates in the south and increasing rates to the north and west, reflecting a shift in exhumation from the southern, internal massifs and nappes to the northern external massifs (Fugenschuh & Schmid 2003). This observation has focused attention on the external massifs as the place to resolve the timing of any acceleration in late-stage erosion rates in the Alps.

The Mt. Blanc external crystalline massif has seen a number of thermochronometry studies (Fugenschuh & Schmid 2003, Glotzbach et al. 2008, Leloup et al. 2005, Seward & Mancktelow 1994). The most recent and complete study was by Glotzbach et al. (2008), who used apatite fission-track ages and apatite (U-Th)/He ages from the surface and the Mt. Blanc tunnel to construct age-elevation relationships. They interpreted these data as indicating rapid cooling between ~8 and 6.5 Mya, slow cooling between 6.5 and 3 Mya, and again rapid cooling since ~3 Mya. This interpretation is based primarily on the fission-track data, which do indicate high rates of exhumation (1 to 2.5 mm year^{-1}) over the earlier time window. However, the slowing of exhumation 6.5 Mya is based on few data and is not supported by the (U-Th)/He data, which also show a linear age-elevation relationship that implies an exhumation rate of ~1 mm year^{-1} over the time window of 5.5 Mya to at least 4 Mya. Taken together, these data seem consistent with an exhumation rate of more than 1 mm year^{-1} from ~8 to 4 Mya. Zircon fission-track data indicate lower cooling rates between 20 and 12 Mya. The apatite He ages in the Mt. Blanc tunnel and in the adjacent deep valleys are all younger than the age-elevation trend, with average ages of approximately 1.6 Ma. These ages could be interpreted as indicating lower cooling rates between 4 and 1.6 Mya, in which case the rates between 1.6 Mya and the present would likely need to be higher. However, these are also the samples most likely to be affected by recent changes in relief, which complicates such interpretations.

The Argentera Massif in the southwestern Alps is the least likely to have seen recent tectonic activity given its location in the westernmost Alps, which presently appear to have no convergent tectonics (Sue et al. 2007) but still exhibit apatite fission-track ages as young as 2.4 Ma (Bigot-Cormier et al. 2006, Bogdanoff et al. 2000). Bigot-Cormier et al. (2006) established two age-elevation relationships that show differences in cooling history between a northeast and a southwest tectonic block within the massif, but both regions show exhumation at a rate of 0.2 km Ma^{-1} between 12 and ~6 Mya. The northeast block accelerates with rapid exhumation (more than 1.0 km Ma^{-1}) starting somewhere between 7 and 5 Mya, resolvable until only 4 Mya. In contrast, the southwest block shows no acceleration until between 4.5 and 2.5 Mya, but there are few data to control this event. As in other external massifs, zircon fission-track ages indicate cooling ~22 Mya, indicating that tectonic activity initiated much earlier than the accelerated cooling recorded by the apatite.

The Aar Massif, the largest of the external massifs, has been the subject of a number of fission-track (Michalski & Soom 1990, Seward & Mancktelow 1994, Wagner et al. 1977) or combined

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fission-track and (U-Th)/He thermochronometry studies (Aramowicz et al. 2007, Glotzbach et al. 2010, Reinecker et al. 2008, Vernon et al. 2009b). These latter studies have focused on the most recent exhumation history. Reinecker et al. (2008) studied the westernmost Aar Massif with thermochronometry samples from the surface and the Lötschberg Tunnel, which crosses the Gastern and Aar massifs. They interpreted a younging trend in age toward the south as evidence for extensional unroofing and footwall tilting of the Aar to the north. Removing this effect, they found that all fission-track ages were consistent with a constant rate of exhumation of 0.5 km Ma^{-1} over the time interval of 10 to 3.5 Mya. However, the youngest ages in this region are those that are arguably affected by extensional exhumation, so if these are removed, the constant exhumation rate is inferred to be valid until ~4.5 Mya. Postclosure exhumation can be calculated by assuming a depth to the closure temperature, and it suggests a similar or possibly slightly higher exhumation rate between 4.5 Mya and the present, again ignoring the tectonically exhumed samples.

In a fission-track study in the eastern central Aar and adjacent Gotthard Massif, Glotzbach et al. (2010) obtained a result similar to Reinecker et al. (2008). They found cooling rates of 0.5 km Ma^{-1} between 9 and 6 Mya, although there was considerable scatter in the age-elevation relationship that permits variations in rate during this interval. This rate is consistent with earlier apatite fission-track work (Wagner et al. 1977). Track-length modeling by Glotzbach et al. (2010) suggests a slightly higher cooling rate between ~10 and 7 Mya relative to subsequent rates, but again the data do not require a change in cooling rate over the past 10 Ma.

Aramowicz et al. (2007) obtained apatite and zircon (U-Th)/He ages from the same region but reached a different conclusion. They found rapid cooling that corresponds to rates of more than 1 km Ma^{-1} over the time interval of 7 to 5.5 Mya. Subsequent cooling rates to the present day are not well resolved in time but must have been slower, on average, with rates of ~0.3 to 0.5 km Ma^{-1}. Vernon et al. (2009b) obtained a similar result for the central Aar Massif. They also found evidence of rapid cooling, but earlier, with a rate of ~1.0 km Ma^{-1} between 9 and 7 Mya as inferred from the elevation distribution of apatite fission-track ages. They also inferred a subsequent lower exhumation rate of ~0.5 km Ma^{-1} from approximately 8 to 4 Mya based on apatite (U-Th)/He ages, although the age-range overlap means these data are not internally consistent. Nonetheless, the rates of 1 km Ma^{-1} could not be sustained until the present day, so some slowing in cooling rate must have occurred at some point after 7 Mya.

The exhumation timing of the external massifs is summarized in **Figure 7**. The spatially averaged exhumation rate for the Western Alps from Vernon et al. (2008) is shown with some select cooling paths from the same source. The colored bars show the timing of rapid exhumation of various external massifs; the actual rate of exhumation is not shown. Vernon et al. (2008) found an overall increase in exhumation rate over the past 5 Ma but could not easily determine where this increase occurred. They found high erosion rates in a band across the external Alps between 5 Mya and the present (as in **Figure 6**), but their analysis showed large variations in space and time. It is difficult to establish how much of this result is an artifact of the analysis method because the researchers did not track resolution errors. Comparison of the regional study of Vernon et al. (2008) with the site-specific studies (**Figure 7**) illustrates this problem through large inconsistencies among the various studies. In particular, although the Alps as a whole show an increase in erosion rate ~5 Mya, individual regions show accelerations at different times.

DEVELOPMENT OF HIGH RELIEF

Thermochronometric data also have the potential to directly measure an increase in relief (Braun 2002a,b; Herman et al. 2007; House et al. 2001). However, only large changes in relief can be

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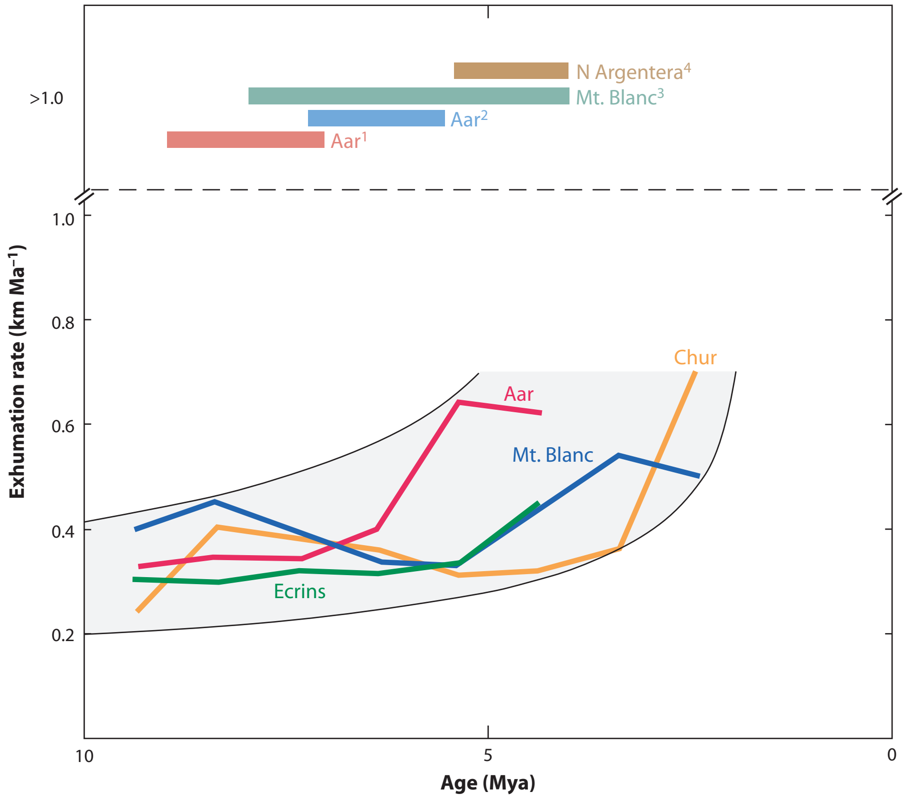

Figure 7

Exhumation rates inferred from low-temperature thermochronometry data. Gray area is average for the Western Alps from Vernon et al. (2008). Colored lines are specific regions isolated by Vernon et al. (2008). Colored bars are times of fast exhumation identified by 1, Vernon et al. (2009b); 2, Aramowicz et al. (2007); 3, Glotzbach et al. (2008); 4, Bigot-Cormier et al. (2006). Although there is a general increase in exhumation ∼5 Mya, the variation in timing is large.

measured, and even then, resolving such signals may depend on new techniques such as ^{4}He/^{3}He measurement in apatite (Shuster et al. 2005). Some thermochronometry studies claim to have detected a signal, or at least to have found a better explanation for their data with a recent increase in relief (Glotzbach et al. 2008, Vernon et al. 2009b), but these studies have not demonstrated that an increase in relief is required to explain thermochronometric ages.

A better argument was forwarded by Haeuselmann et al. (2007), who determined burial ages for detrital quartz in a cave system in central Switzerland. By using cosmogenic ^{10}Be and ^{26}Al to date the burial of quartz sand washed into the cave, they could establish the level of the valley adjacent to the cave system. They demonstrated that the base level for the cave hydrologic system dropped rapidly starting between 1 and 0.8 Mya, which they relate to downcutting of the Aar River system. This argues for major, glacial incision of the Alpine valleys well after the onset of Northern Hemisphere glaciation, although an age of 0.87 Ma corresponds well with marine isotope stage 22, which has been proposed, based on sedimentological evidence, as the time for onset of major glaciation in the Alps (Mancin et al. 2009, Muttoni et al. 2003).

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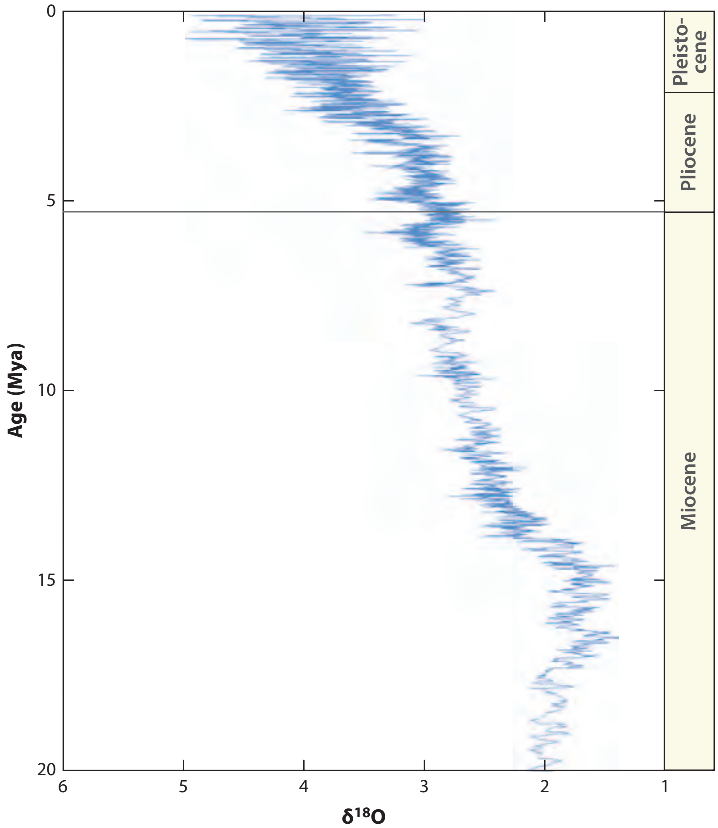

Figure 8

Oxygen proxy record of global climate change based on deep-ocean record. Data from Zachos et al. (2001).

CLIMATE HISTORY OF THE ALPS SINCE THE LATE MIOCENE

The late Miocene to the present marks the culmination of the global Cenozoic cooling trend that includes Northern Hemisphere glaciation. Globally, both cooling and increased variability characterize the past 15 Ma (Zachos et al. 2001) (**Figure 8**). Since the mid-Miocene climatic optimum, the Miocene was characterized by relatively cool conditions with periodic ice-volume fluctuations thought to be in both the Southern Hemisphere and the Northern Hemisphere. The early Pliocene represents a short hiatus in the cooling trend, with warmer conditions relative to the late Miocene and the later Plio-Pleistocene (Raymo et al. 1996). It has been argued that this warming resulted from closure of the Panama gateway and subsequent strengthening of the North Atlantic gyre (Haug & Tiedemann 1998, Haug et al. 2001). Since the mid-Pliocene, the global climate has become progressively colder, and more variable with increased amplitude in the orbitally driven cyclicity (Zachos et al. 2001). Both cooling and cyclicity increased dramatically 2.7 Mya, coincident with major Northern Hemisphere glaciation. In the mid-Pleistocene, between ~1 and 0.8 Mya, there was an additional major climate change with a power shift from the 40,000-year period to the 100,000-year period, which has dominated global climate since (Imbrie et al. 1993, Raymo et al. 1997).

The problem, as always, is how to translate these global records into a regional record—particularly with respect to precipitation rate—that is relevant for the erosional record of the Alps.

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One solution is to find correlations between global and regional data. For example, the paleofloral record of central Europe has been used to reconstruct temperature and precipitation conditions for the lowlands north of the Alps (Donders et al. 2009; Mosbrugger et al. 2005; Utescher et al. 2000, 2009). These proxy data show a decrease with temperature and an increase in seasonality over the past 10 Ma, but they show no dramatic change in precipitation that could be responsible for the increased erosion over the past 5 Ma. In fact, one of the most dramatic changes in the northern European record is an inferred decrease in precipitation in the Zanclean (Utescher et al. 2000). However, the fossil and pollen data come primarily from the northern European platform basins, so, although relevant for the continent, they could easily miss effects local to the Alps.

The southern European and Mediterranean palynological record also offers no strong evidence for climatic change that correlates with the sediment-yield data, although the overall pattern of cooling and increased variability, particularly in the Pleistocene, is evident. In the earlier record, the MSC has been the focus of several high-resolution paleoclimate studies, in part to address the question of the influence of aridity and sea level on isolation and desiccation of the Mediterranean (Bertini 2006, Fauquette et al. 2006, Pierre et al. 2006, Warny et al. 2003). There are no dramatic changes in flora across the MSC, although, again, studies are not specific to the Alps but instead cover much of the Mediterranean with its highly diverse flora, which makes interpretation difficult. A major flora shift internal to the MSC occurred 5.5 Mya, which corresponds to the onset of the Lago Mare environmental conditions. This shows up primarily as an increase in Pinaceae occurrence (Bertini 2006, Iaccarino et al. 2008). The relative frequency of Pinaceae pollen is generally not considered a reliable climate indicator because of its transportability; rather, it is interpreted as a response to shoreline distance and preferential preservation. However, during the late MSC, there was no connection between the Mediterranean and the global ocean, so any change in sea level was in response to the hydrologic balance in the Mediterranean as opposed to global eustatic change. Therefore, Pinaceae is responding either to higher local precipitation or to the higher water levels in the Mediterranean, which also is a reflection of higher precipitation. In either case, Pinaceae should be considered a climate proxy. Water levels in the Mediterranean were high during deposition of the Lago Mare sediments (Bassetti et al. 2004, Keogh & Butler 1999), adding support to the inference that precipitation increased 5.5 Mya. That this sedimentological change also corresponds perfectly to the global warming event at the end of the last Miocene glacial period (Hodell et al. 2001, van der Laan et al. 2006, Vidal et al. 2002) also supports this argument.

The early Pliocene continued to be wet and warm, with precipitation rates two to three times modern rates as estimated by palynological methods (Fauquette et al. 1998). If the Pliocene warm period arose from closure of the Panama seaway, it is likely that Europe was strongly affected by enhanced North Atlantic circulation; this is consistent with the regional climate data. Closure of the gateway was progressive, and final closure did not occur until 2.7 Mya [see review by Molnar (2008)]. Furthermore, there was still free exchange of water between the ocean basins 4.6 Mya (Groeneveld et al. 2008). The connection to deep-water circulation in the Atlantic is still nebulous, but the changes to Caribbean oceanography thus suggest that closure postdates the start of the Pliocene (Haug & Tiedemann 1998). Therefore, closure does not correlate well with the initial increase in sediment flux from the Alps.

By mid-Pliocene, the global cooling is evident also in the Mediterranean. In addition, when Northern Hemisphere glaciation fully set in 2.7 Mya, the regional climate shows both the cooling and the fluctuations between glacial and interglacial climates (Fauquette et al. 1998). Although, as noted above, Alpine glaciation, or at least glacial erosion, was likely only minor until the mid-Pleistocene transition that occurred 0.87 Mya (Muttoni et al. 2003).

DISCUSSION

The observations presented above draw a complicated picture of major changes in Alpine erosion rates through the late Neogene, and it remains to be seen whether a consistent story emerges. A number of provocative questions remain. Foremost of these is, why? Is this a response to a changing climate, or is there a tectonic component to the system? If it is a climate response, what aspects of climate change are important to erosion rates and when were these initiated? The second set of questions center on the sediment source area. Can we identify where the enhanced sediment production originated? Is there a spatial pattern to enhanced exhumation that might fingerprint the driving mechanism, regardless of whether it is tectonic or climatic? These remain unanswered questions, and it is clear that better space and time resolution of the relevant data is essential. Still, we can begin to put together a picture of the events of the late Neogene.

In the late Messinian, the Western Alps were active, given the evidence of convergent deformation on the southern (Schumacher et al. 1996) and northern (Becker 2000, Bolliger et al. 1993) margins of the range. In fact, given that both margins were active, the Alps were expanding at this time in a constructive state. Interior to the orogen, the external massifs either were exhuming rapidly or were in the process of slowing exhumation; this likely indicates that they were tectonically active but perhaps at the end of their active phase. The climate, at least south of the Alps, was dry enough to create the negative hydrologic balance in the Mediterranean that made the MSC possible. Near the end of the Messinian, in the last stages of the MSC, the sediment flux from the Alps to the Adriatic Basin and Po Basin dramatically increased at the same time that the water budget for the Mediterranean shifted toward a more positive precipitation/evaporation ratio. This occurred precisely at the end of the last Miocene glacial period, suggesting a global-climate connection (Vidal et al. 2002). The correlation with the end of deformation in the Lombardic thrust belt suggests a causal relationship, which could reflect the erosional destruction of the topographic gradient in the Southern Alps in response to the wetter climate and the deep incision of MSC valleys into the Southern Alps. The additional mechanism of the sediment load deposited on top of the previously active thrust sheets in the Po Plain could also shut down deformation, but this is effectively the same mechanism of changing surface slope by surface mass transport. Corresponding events to the north include the end of deformation in the Jura, uplift and erosion of the NAFB, and increased sediment load to the Rhine and Rhone, but these events are not dated well enough to be conclusively correlated with those in the south.

The warm climate of the early Pliocene was more erosive than the preceding Miocene climate, and the cold, more variable climate of the Pleistocene was even more erosive, given that each of these times is associated with an additional increase in sediment yield. However, the existing evidence supports only minor glaciation of the Alps in the early Pleistocene, so this must have been an efficient, but fluvial, geomorphic system. At the mid-Pleistocene climate shift 0.87 Mya, glaciation impacted the Alps more seriously; this is associated with yet another increase in net erosional flux. The landscape may have been significantly modified at this time, with deepening and flattening of large valleys and some increase in local relief, but the preexisting landscape was likely to have had significant fluvial relief, so it is not clear that relief change was as important as the increase in net erosion rate. It is somewhat troubling that every change in climate, regardless of whether it is warming or cooling, leads to higher erosion rates, but that seems to be the observation through the Plio-Pleistocene. Alternatively, it is not the mean temperature that matters, but rather the climate variability and the rate of change of climate, both of which increase toward the present day. Both active and passive tectonic models predict a progressive decrease in sediment yield in response to a stepwise increase in erosional efficiency, so it is clear that we are not dealing with a single change, but rather a progressively more erosive climate over the past 5.5 Ma.

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The paleoclimate inferred by flora studies shows changes at key times, but these changes are subtle in comparison with the changes in sediment yield. One explanation could be differential response of erosional processes to precipitation rate. Fluvial erosion rates scale nonlinearly with discharge and thus with precipitation, so that erosion rates potentially scale not with the mean precipitation rate but instead with some characteristic storm size (Lague et al. 2005). Thus a small increase in mean precipitation can lead to a disproportionate increase in the number of highly erosive storms.

The present-day erosion rates are well constrained and reasonably consistent across different measurement techniques. Sediment yield based on sedimentary-basin volume assessments (Kuhlemann 2000), postglacial lake filling (Hinderer 2001), basement fission-track age modeling (Vernon et al. 2008), and detrital ^{10}Be studies (Wittmann et al. 2007) all yield an estimated erosion rate of $\sim$0.6 km Ma^{-1}. There is significant uncertainty attached to each of these measurements, but it is comforting that they agree so closely. It is also interesting that the pattern of geodetically derived, modern rock-uplift data also matches the pattern and magnitude of long-term erosion rates. However, this correspondence does not help differentiate between models because most models predict a ratio of erosion rate to rock-uplift rate between 0.8 and 1.0—precisely the range permitted by uncertainty in the data correlation. In addition, the time-dependent uplift in response to ice retreat must be carefully assessed, and this limits the usefulness of the modern data (Stocchi et al. 2009).

Matching sediment source area to sediment flux remains highly problematic. Although there does appear to be an overall increase in integrated erosion rate inferred from thermochronometry and a general shift of erosion from the southern, internal Alps to the northern, external Alps, it is difficult to identify specific regions that experienced accelerated exhumation at the same time that sediment yield increased. That detrital fission-track studies do not detect any change in exhumation rate is consistent with the idea that the regions experiencing accelerated exhumation are not the regions with the youngest zircon fission-track basement ages. Specific studies of the external crystalline massifs indicate either no acceleration in exhumation over the past 5 to 10 Ma or an acceleration between $\sim$9 and 6 Mya—too early to provide an explanation for the sediment yield increase. These findings likely suggest a response to tectonic uplift of individual massifs. The uncertainty in timing reflects different analysis techniques; studies not designed to resolve late cooling; or broad, regional studies that cannot easily resolve local processes. Therefore, this problem awaits more detailed research.

The question of an active or passive tectonic response to a more erosive climate remains unresolved. The correlation in time between cessation of mountain front deformation and the increase in sediment yield is strong evidence for an active response. The active tectonic response predicts sustained higher exhumation in the orogen interior, but this will be nearly impossible to resolve given the few million years of observation time and the 20-Ma response time of the Alps.

In addition, coincidence cannot be ruled out, and one can envision more complex alternative models. It is possible that the event 5.5–5 Mya is the cessation of tectonic convergence in the Western Alps. Most models—and intuition—maintain that the end of tectonic convergence should be associated with a decrease in sediment yield, but if the topographic load is regionally supported, the shift from constructive to destructive state leads to regional rock uplift (**Figure 4*b***). The peripheral regions of the orogen, including the foreland basins, would uplift and erode. If these regions erode more easily than the orogenic core, as is almost certainly the case, there could be a net increase in sediment yield. Normally, this increase would be short lived with progressively decreasing sediment flux, but if this tectonic event were closely followed by the climatic events of the Plio-Pleistocene as well as the erosional impact of the MSC, these signals could be constructively mixed. These remain open questions.

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CONCLUSIONS

In studies of orogenic mechanics, erosional destruction of topography, and impacts of climate change on mountain-belt erosion and tectonics, the Alps provide a prominent and well-studied example. It seems clear that erosion rates increased in the latest Miocene and have continued to accelerate toward the present day. Given the observation that tectonic processes have slowed or stopped in the Western Alps over this same time period, we are left with climatic processes as the prime candidate for the driving mechanism. Beyond these basic observations, there are mostly questions regarding specific mechanisms, feedback, and timing. The problems have been well defined by studies of sediment budgets, exhumation rates, and erosion rates, but clear solutions have not emerged. New techniques, particularly in low-temperature thermochronometry, have opened new opportunities to resolve changes in erosion rates and patterns over the past 10 Ma. Ultimately, the questions of the control on the past 5 to 10 Ma of erosion, through the dramatic climate change of the Plio-Pleistocene, are important enough that the Alps will continue to be an important testing ground for new techniques, methods, models, and ideas.

DISCLOSURE STATEMENT

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Annual Review of Earth and Planetary Sciences

Volume 38, 2010

Contents

Frontispiece
Ikuo Kushiro xiv

Toward the Development of "Magmatology"
Ikuo Kushiro 1

Nature and Climate Effects of Individual Tropospheric Aerosol Particles
Mihály Pósfai and Peter R. Buseck 17

The Hellenic Subduction System: High-Pressure Metamorphism, Exhumation, Normal Faulting, and Large-Scale Extension
Uwe Ring, Johannes Glodny, Thomas Will, and Stuart Thomson 45

Orographic Controls on Climate and Paleoclimate of Asia: Thermal and Mechanical Roles for the Tibetan Plateau
Peter Molnar, William R. Boos, and David S. Battisti 77

Lessons Learned from the 2004 Sumatra-Andaman Megathrust Rupture
Peter Shearer and Roland Bürgmann 103

Oceanic Island Basalts and Mantle Plumes: The Geochemical Perspective
William M. White 133

Isoscapes: Spatial Pattern in Isotopic Biogeochemistry
Gabriel J. Bowen 161

The Origin(s) of Whales
Mark D. Uhen 189

Frictional Melting Processes in Planetary Materials: From Hypervelocity Impact to Earthquakes
John G. Spray 221

The Late Devonian Gogo Formation Lägerstatte of Western Australia: Exceptional Early Vertebrate Preservation and Diversity
John A. Long and Kate Trinajstic 255

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Booming Sand Dunes
Melany L. Hunt and Nathalie M. Vriend 281

The Formation of Martian River Valleys by Impacts
Owen B. Toon, Teresa Segura, and Kevin Zahnle 303

The Miocene-to-Present Kinematic Evolution of the Eastern Mediterranean and Middle East and Its Implications for Dynamics
Xavier Le Pichon and Corné Kreemer 323

Oblique, High-Angle, Listric-Reverse Faulting and Associated Development of Strain: The Wenchuan Earthquake of May 12, 2008, Sichuan, China
Pei-Zhen Zhang, Xue-ze Wen, Zheng-Kang Shen, and Jiu-hui Chen 353

Composition, Structure, Dynamics, and Evolution of Saturn's Rings
Larry W. Esposito 383

Late Neogene Erosion of the Alps: A Climate Driver?
Sean D. Willett 411

Length and Timescales of Rift Faulting and Magma Intrusion: The Afar Rifting Cycle from 2005 to Present
Cynthia Ebinger, Atalay Ayele, Derek Keir, Julie Rowland, Gezahegn Yirgu, Tim Wright, Manahloh Belachew, and Ian Hamling 439

Glacial Earthquakes in Greenland and Antarctica
Meredith Nettles and Göran Ekström 467

Forming Planetesimals in Solar and Extrasolar Nebulae
E. Chiang and A.N. Youdin 493

Placoderms (Armored Fish): Dominant Vertebrates of the Devonian Period
Gavin C. Young 523

The Lithosphere-Asthenosphere Boundary
Karen M. Fischer, Heather A. Ford, David L. Abt, and Catherine A. Rychert 551

Indexes

Cumulative Index of Contributing Authors, Volumes 28–38 577

Cumulative Index of Chapter Titles, Volumes 28–38 581

Errata

An online log of corrections to *Annual Review of Earth and Planetary Sciences* articles may be found at http://earth.annualreviews.org

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TABLE OF CONTENTS:

- *What Is Statistics?* Stephen E. Fienberg
- *A Systematic Statistical Approach to Evaluating Evidence from Observational Studies,* David Madigan, Paul E. Stang, Jesse A. Berlin, Martijn Schuemie, J. Marc Overhage, Marc A. Suchard, Bill Dumouchel, Abraham G. Hartzema, Patrick B. Ryan
- *The Role of Statistics in the Discovery of a Higgs Boson,* David A. van Dyk
- *Brain Imaging Analysis,* F. DuBois Bowman
- *Statistics and Climate,* Peter Guttorp
- *Climate Simulators and Climate Projections,* Jonathan Rougier, Michael Goldstein
- *Probabilistic Forecasting,* Tilmann Gneiting, Matthias Katzfuss
- *Bayesian Computational Tools,* Christian P. Robert
- *Bayesian Computation Via Markov Chain Monte Carlo,* Radu V. Craiu, Jeffrey S. Rosenthal
- *Build, Compute, Critique, Repeat: Data Analysis with Latent Variable Models,* David M. Blei
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- *Event History Analysis,* Niels Keiding
- *Statistical Evaluation of Forensic DNA Profile Evidence,* Christopher D. Steele, David J. Balding
- *Using League Table Rankings in Public Policy Formation: Statistical Issues,* Harvey Goldstein
- *Statistical Ecology,* Ruth King
- *Estimating the Number of Species in Microbial Diversity Studies,* John Bunge, Amy Willis, Fiona Walsh
- *Dynamic Treatment Regimes,* Bibhas Chakraborty, Susan A. Murphy
- *Statistics and Related Topics in Single-Molecule Biophysics,* Hong Qian, S.C. Kou
- *Statistics and Quantitative Risk Management for Banking and Insurance,* Paul Embrechts, Marius Hofert

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