

Surface uplift, uplift of rocks, and exhumation of rocks

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ABSTRACT

Uplift of the surface of mountain belts requires forces that are comparable in magnitude to those associated with plate motion, and therefore determination of rates of surface uplift could provide important information on the dynamics of mountain ranges. Rates of uplift of the surfaces of mountain ranges have not, however, been quantified sufficiently well that they provide useful constraints on those processes. Many reports of surface uplift in mountain ranges are based on mistaking exhumation of rocks or uplift of rocks for surface uplift, and provide no information whatsoever on the rates of surface uplift.

UPLIFT AND EXHUMATION

What is Uplifted with Respect to What?

The purposes of this paper are (1) to make clear why, in tectonics, the rate of uplift of the surface with respect to the geoid is the important uplift rate to measure and (2) to draw attention to the origins of the confusion surrounding some attempts to make such measurements.

The word “uplift” refers to displacement in the direction opposite to the gravity vector. A displacement is only defined when both the object displaced and the frame of reference are specified. Confusion has arisen frequently in the structural geology and tectonics literature because either the object or the frame of reference is not specified; sometimes neither is.

The objects most commonly considered when the word “uplift” is used in a tectonic context are rocks and parts of Earth’s surface. The frame of reference used is usually attached either to Earth’s surface or to the geoid. There are, therefore, three kinds of displacement to which the terms “uplift” and “uplift rate” are applied.

1. Displacement of Earth’s surface with respect to the geoid: we refer to this displacement as “surface uplift.” It is important to recognize that a spot height is rarely of interest in tectonic problems. The scale of tectonic processes controlling mountain building is at least that of the crustal thickness (≥ 30 km); when we refer to “the surface” we mean, therefore, the interface between rock and air (or water) over a region whose area is at least 10^3 – 10^4 km². When we refer to “surface height” and “surface uplift” we have in mind quantities obtained by averaging elevation and changes in elevation over surface areas of that size. (For example, whereas the peak of Mt. Everest is nearly 9 km high, the surface height averaged over 1° squares in latitude and longitude in the neighborhood of Mt. Everest is about 5 km; the latter quantity is the important one in our discussion: see Frank, 1972; Molnar and Lyon-Caen, 1988; and the next section, below.)

2. Displacement of rocks with respect to the geoid: we refer to this displacement as “uplift of rocks.”

3. Displacement of rocks with respect to the surface: we refer to this displacement as “exhumation.” The rate of exhumation is simply the rate of erosion or the rate of removal of overburden by tectonic processes.

Because uplift is defined in terms of movement in the opposite direction to the force of gravity, it is natural to choose a frame of reference attached to an equipotential surface of gravity. One such surface is the present geoid. With such a frame of reference, a given displacement involves a quantifiable amount of work against gravity. In practice, of

course, no serious error would be introduced if the phrase “mean sea level” were substituted for “geoid,” provided that account were taken of eustatic changes in sea level. We use the geoid as our frame of reference for discussing uplift.

It is evident that the three displacements are related:

$$\text{surface uplift} = \text{uplift of rock} - \text{exhumation}. \quad (1)$$

All three displacements cannot be equal, except in the singularly uninteresting case where all three are zero. The assumption (often made implicitly) that, for example, uplift of rock and surface uplift are equal is equivalent to the assumption that the exhumation is zero—often a difficult assumption to justify.

Driving Forces for Uplift

The raised surfaces of mountain ranges result largely from thickening of the continental crust. The horizontal shortening that causes that thickening takes place at plate tectonic rates, a few centimetres a year. Thus, to take a specific example, India has been converging with Asia at ~ 50 mm·yr⁻¹ for about the past 50 m.y., and the demonstration that the surface of the Tibetan plateau rose from below sea level to its present height of 5 km at a rate that was roughly constant (about 0.1 mm·yr⁻¹) over this interval of time would hold no surprises. A stimulus for this paper is that much attention has been paid in recent years to the notion that the surfaces of mountain ranges may have increased in height at a rate that cannot be accounted for by crustal thickening alone.

If the study of tectonics is to be more than a merely descriptive pastime, it must involve consideration of the forces responsible for tectonic activity. All tectonic processes are driven by the force of gravity in the presence of differences in density. At the largest scale, this is manifested by circulation of the mantle, of which plate motion is a part. The creation, cooling, and subduction of the oceanic plates produce lateral density variations near the surface. These density differences are generally compensated isostatically, and the consequent lateral variations in the vertically averaged normal stress on vertical planes are often referred to as the “driving force” (per unit length) associated with plate motion (e.g., Richter and McKenzie, 1978). The “driving force” exerted by one column of lithosphere upon another is physically identical to the difference in potential energy per unit surface area between the two columns (e.g., Molnar and Lyon-Caen, 1988).

The dynamics of mountain ranges are governed by differences in the gravitational potential energy of entire columns of the continental lithosphere (e.g., see Artyushkov, 1973; England and Houseman, 1989; Frank, 1972; Molnar and Lyon-Caen, 1988). If changes in surface height represent isostatically compensated changes in crustal thickness, the rate of change of potential energy per unit surface area, dPE/dt , is proportional to the rate of increase of average surface height (dh/dt):

$$\frac{dPE}{dt} = C \frac{dh}{dt}, \quad (2)$$

where the constant, C , is proportional to crustal thickness, t is time, and h

is average surface height. This result follows directly from the linear dependence of the surface height on the crustal thickness and the dependence of the potential energy difference, or "driving force," on the square of the crustal thickness (e.g., Artyushkov, 1973; England and Houseman, 1989; Frank, 1972; Molnar and Lyon-Caen, 1988). The magnitude of the constant of proportionality, C , depends on the change in thickness of the mantle part of the lithosphere that accompanies a given change in the crustal thickness. If the mantle part of the lithosphere thickens in proportion to the crust, the change in potential energy can be small, and might, in some cases, be negative (England and Houseman, 1989; Fleitout and Froidevaux, 1982). The highest rates of working against gravity are required when the thickness of the crust, but not of the lithosphere, changes. Under these circumstances, $C \approx 1 - 2 \times 10^9 \text{ N}\cdot\text{m}^{-2}$, for crustal thicknesses ranging from 35 to 70 km (England and Houseman, 1989, Appendix A) and the rate of working associated with a surface uplift rate of $1 \text{ mm}\cdot\text{yr}^{-1}$ is about $35 \text{ to } 70 \text{ mW}\cdot\text{m}^{-2}$.

A major concern of this paper is with surface uplift that is *not* associated with changes in crustal thickness. The most plausible mechanism causing such uplift is the application of normal stresses to the base of the lithosphere by convection in the mantle (e.g., Houseman and England, 1986). England and Houseman (1988) suggested that the convective instability of the thickened continental lithosphere (Houseman et al., 1981) could lead to the rapid thinning of the mantle part of the lithosphere beneath mountain ranges and a rapid increase in surface height, without change in crustal thickness. Uplift of this nature would require a rate of working of about $100 \text{ mW}\cdot\text{m}^{-2}$ for an uplift rate of $1 \text{ mm}\cdot\text{yr}^{-1}$.

The shear stresses resisting motion on major thrust faults are of probably about 100 MPa (Molnar and England, 1990). The rate of working associated with slip on a fault is the product of the shear stress and the rate of slip. For slip at $20 \text{ mm}\cdot\text{yr}^{-1}$ on a fault along which the shear stress is 100 MPa, power is dissipated at a rate of $60 \text{ mW}\cdot\text{m}^{-2}$; thus surface uplift at a rate of about $1 \text{ mm}\cdot\text{yr}^{-1}$ may represent a significant fraction of the work done in creating a mountain range.

Isostatically compensated exhumation, acting alone, reduces the crustal thickness and lowers the surface height. Whether such exhumation results from tectonic activity or from erosion, the gravitational potential energy of the lithosphere decreases. The change in potential energy of a column of lithosphere associated with uplift of the rocks within it cannot be quantified unless the rates of both exhumation and change of surface height are known.

Thus, the only uplift that is associated with a quantifiable amount of work done against gravity is surface uplift. For this reason, and because that amount of work is large, surface uplift is the displacement that one should measure in order to obtain information about the tectonic forces acting in mountain belts.

MEASURING RATES OF UPLIFT AND EXHUMATION

Uplift

Measuring the uplift of a point requires measuring its change in elevation. Measuring surface uplift over a finite area requires measuring, or estimating, the change in elevation of *all* points on the surface of that area. Only where the rate of erosion is zero do measurements of displacements of points on the surface reflect regional surface displacements.

Determination of the rate of surface uplift depends upon the preservation of a sequence of rocks deposited during that uplift, and upon those rocks containing evidence of surface elevation at the times of their deposition. When the uplift takes place far above sea level, these conditions are not easily met; erosion is more common than deposition in mountainous regions, and few features preserved in terrestrial sediments reflect directly the surface height at the time of deposition.

In some cases, sediments or sedimentary rocks containing evidence of

their elevation at the time of deposition have been uplifted over large areas with little apparent accompanying erosion, so that it is reasonable to regard the uplift of these rocks as a measure of the surface uplift. For example, the uplift of coastal terraces provides valuable information on rates of surface movement in tectonically active areas (e.g., Lajoie, 1986). In some cases, an observation of uplifted rock provides valuable evidence of surface uplift, even when the erosion of the region is unquantified. For example, the widespread occurrence in southern and western Tibet of Cretaceous marine limestone at a present altitude of 5 km provides the most definitive constraint on the magnitude of Cenozoic surface uplift in that region.

It is often not possible, however, to infer *rates* of surface displacement solely from the present positions of rocks that are far above sea level. For example, the presence of Cretaceous marine limestones in Tibet gives no information on when, nor at what rate, their uplift occurred. In particular, we cannot distinguish between a recent surface uplift and a recent surface subsidence in the region on the basis of the present elevation of those rocks.

A method that offers some promise for determining paleoelevations, and hence surface uplift rates, relies on using paleontology as a paleoclimatic tool. Climates vary markedly with elevation and paleontological observations can yield information on paleoclimate (e.g., Wolfe, 1971, 1978, 1979). This approach must be used with caution because, in addition to depending on elevation, climates also vary with latitude, and have changed, both regionally and globally, over geologic time. The procedure involves, therefore, first inferring changes in climate in an area of interest from changes with time of floral assemblages that are known to be sensitive to temperature and rainfall (e.g., Meyer, 1986; Wolfe, 1978; Wolfe and Schorn, 1989). Secular climatic changes occurring over the same interval are detected by observation of organisms that lived at sea level in the same general region and at the same time. Local climatic changes are then corrected for regional secular changes before being used to infer elevation changes in the area of interest (e.g., Meyer, 1986).

Exhumation

The magnitude of the exhumation a rock has undergone may be inferred from geobarometry. Geothermometry, coupled with an estimate of the ambient geothermal gradient, may also yield bounds on the exhumation of rocks. Geobarometers and geothermometers contain no information about time, so these techniques alone cannot yield a rate of exhumation.

The cooling age of a mineral should record the time at which it became cool enough to retain the daughter products or the fission tracks produced by the decay of radioactive isotopes. The different temperatures at which minerals close with respect to the different products of radioactive decay allow determination of a set of times when a rock cooled below various blocking temperatures. With the assumption of a thermal profile, the rate of exhumation can be calculated from such measurements. Inferences of exhumation rates from fission tracks and radiometric ages are one of the major advances in geochronology over the past 20 yr (e.g., Clark and Jäger, 1969).

HAVE THE SURFACES OF MOUNTAIN RANGES RISEN RAPIDLY?

Many *reported* rates of uplift in mountain ranges, treated implicitly or explicitly as rates of surface uplift, are too rapid to be accounted for by known rates of crustal thickening; e.g., the rates of $4\text{--}5 \text{ mm}\cdot\text{yr}^{-1}$ quoted by Copeland et al. (1987) for southern Tibet, of $1\text{--}100 \text{ mm}\cdot\text{yr}^{-1}$ by Xu (1981) for the whole of Tibet, and of $0.7 \text{ mm}\cdot\text{yr}^{-1}$ by Benjamin et al. (1987) for the Andes. Furthermore, there are regions in which the reported surface uplift apparently postdates tectonic activity, and others in which

the surface heights are believed to have been increasing at a time when the tectonic activity suggests crustal thinning.

If uplift were demonstrated to occur at rates faster than can be explained by crustal thickening, that demonstration would provide important information on the processes occurring in the region. However, most reports of anomalous rates of surface uplift are based on misconception, even if such uplift may in fact have occurred. These misconceptions abound in the structural geology and tectonics literature, but are less common among geodesists, sedimentologists, and geomorphologists who are accustomed to use the geoid as a reference level. In the subsections that follow we cite papers providing individual examples of the mistakes that we discuss, so that the reader may recognize such mistakes when they are encountered elsewhere. It is not our intention to provide an exhaustive list of papers containing such mistakes, nor to imply that the papers we cite are less than excellent in other respects.

Mistaking Uplift of Rocks for Surface Uplift

Uplifted rocks can yield reliable estimates of rates of surface uplift when they have not been appreciably dissected by erosion (above). The mistakes we discuss in this section arise principally from overlooking the isostatic adjustment that accompanies erosion. Isostatic adjustment takes place on a time scale of about 10^4 yr (e.g., Cathles, 1975). In consequence, even the most rapid rates of regional erosion (a few millimetres per year; e.g., Bloom, 1978) do not cause the surface to depart from its isostatically balanced level by more than a few tens of metres. (For an incorrect inference, based on exhumation rates, of the rate of isostatic adjustment see Harrison et al. [1989].)

The uplift of *some* of the rocks on the surface of a region is sometimes erroneously used to infer the change in mean surface elevation of the entire region. Consider a portion of the surface whose elevation is governed only by erosion and isostatic adjustment to the erosion. The (purely hypothetical) situation in which erosion occurs at precisely the same rate over the whole surface is not relevant, because such erosion would leave no markers from which uplift could be inferred, and only exhumation could be determined. In any actual region subject to erosion, some portions of the land surface are denuded more rapidly than others because of the influence of rock types, rock structure, drainage patterns, and rainfall. Isostatic response to such differential erosion is regional; i.e., the lithosphere moves upward in response to the *average* removal of load over an area of 10^4 km² or more (see Tsuboi, 1983, p. 203–210). Erosion reduces the thickness of the crust and, in the absence of other influences, the surface therefore moves downward with respect to the geoid as erosion occurs, while individual rocks remaining on or below the surface move upward with respect to the geoid. These individual rocks (or, for example, geodetic markers attached to them) provide the basis for a determination of uplift, but unless precise correction is made for the influence of regional erosion and the isostatic response to it, such a determination applies to the uplift of rocks, not to surface uplift.

Geodetic techniques readily provide present surface heights and re-surveys can yield the displacement of benchmarks on the surface. Displacements of benchmarks, however, do not yield a reliable determination of the change of surface height. Most geodetic networks are not constructed with the intent of determining rates of surface uplift but, even if they were, the only points in a network that could be reoccupied repeatedly for the purposes of determining uplift rates are those not washed away or buried between surveys. Thus in a region that is undergoing isostatically compensated erosion, the markers that remain move upward while the mean surface elevation falls. Geodetic determinations of uplift are, therefore, generally determinations of the uplift of rocks only. An example of the interpretation of rock uplift as surface uplift may be found in Schaer and Jeanrichard's (1974) discussion of leveling across the Swiss Alps.

Although morphological features such as raised river terraces or raised peneplains may tell us about surface uplift, the logic of the preceding paragraphs applies to them also, if they have been dissected. De Sitter (1952) reported a Miocene, and therefore post-tectonic, erosion surface at an elevation of 700–900 m on the north slope of the Pyrenees which he extrapolated to about 3000 m at the center of the range. From this he inferred an increase in surface height of the Pyrenees by about 2000 m since the Miocene epoch. This surface is not continuous, however, but has been highly dissected by streams that created the present juvenile landscape. Hence, the surface uplift is surely less than 2000 m and might be negative. The uplift of rocks that de Sitter (1952) reported might not be tectonically driven at all, but rather the result of denudation and its regional isostatic compensation. Whereas it might be possible to make a correction for the isostatic response to erosion over a large region, it is probably impossible to use the relative elevation or the incision of smaller features such as river terraces to ascertain regional rates of surface uplift.

Mistaking Exhumation for Surface Uplift

Geobarometry and geothermometry provide estimates of exhumation and, combined with geochronology, of exhumation rates (see above). A brief inspection of the uncertainties involved in these measurements indicates that they cannot provide quantitative information on rates of change of surface height.

The lowest temperature associated with a cooling age is 50–100 °C (Faure, 1986); therefore, the amount of exhumation determined from cooling ages is likely to be at least 1–3 km. Exhumation of this magnitude is largely or completely compensated isostatically, so that any associated change in mean surface height would be a small fraction (usually 10%–20%) of the exhumation (e.g., Tsuboi, 1983, Chapter 13). Moreover, calculation of exhumation rates from rates of cooling of rocks requires a knowledge of the temperature profile at the time the cooling occurred. The errors involved in estimating transient temperature profiles at times in the geological past are difficult to quantify and are rarely quoted (see Zeitler, 1985, for a discussion). Those errors are probably never less than 20%, and may often be much more. Estimates of exhumation by the use of geothermometers and geobarometers are also subject to uncertainties (e.g., Powell and Holland, 1988) that exceed in magnitude the changes in surface height that would accompany the exhumation. Thus exhumation rates can tell us nothing about surface uplift rates.

Most reports of surface uplift based on cooling ages assume that an increase in exhumation rate reflects an increase in surface height, perhaps on the grounds that erosion rates generally increase with surface height. This assumption may be correct, but there are two strong arguments against accepting it unquestioned. First, erosion rates depend on other factors than surface height. Second, the exhumation may not be by erosion at all.

Erosion rates certainly do depend strongly on surface height, but they can also vary by an order of magnitude with changes in rainfall (e.g., Bloom, 1978, Table 12.3). In regions that are close to isostatic equilibrium, exhumation is accompanied by a decrease in surface height. Thus an increase in erosion rate may simply reflect a change in climate alone and a more rapid *lowering* of the surface. The marked climatic changes in the last few million years may, for example, be responsible for increased rates of denudation, and the creation of dramatic morphology, without any associated surface uplift (Molnar and England, 1990).

The past 15 yr have brought an appreciation that many mountain ranges may have undergone important phases of extension (e.g., Molnar and Tapponnier, 1975, 1978; Molnar and Chen, 1983; Platt, 1986). Such extension can produce rapid exhumation of rocks and *lowering* of the surface height. Rates of tectonic exhumation, like rates of erosion, cannot be used to determine mean elevation. A quick glance at regions of active

extension should convince anyone that there is no strong correlation between rates of tectonic exhumation and surface height. Continental crust of the Aegean region is largely below sea level, and is extending at about $3 \times 10^{-15} \text{ s}^{-1}$ (Jackson and McKenzie, 1988); the surface of Tibet is at an altitude of 5 km and is extending about ten times more slowly (Molnar and Lyon-Caen, 1989).

Summary

The *uncertainty* in estimates of exhumation is larger than any likely change in surface height accompanying the exhumation, and therefore inferences of exhumation from cooling ages, geobarometry, and geothermometry are useless as indicators of changes in surface height. Furthermore, the common assumption that a measured exhumation rate and its associated rate of surface displacement have the same sign may well be wrong. The mistake of equating an exhumation rate based on rates of cooling of rocks with a rate of surface uplift has been made often, as by Copeland et al. (1987) for Tibet, Zeitler (1985) for the Himalaya, and Benjamin et al. (1987) for the Andes.

Mistaking Changes in Climate for Changes in Surface Height

Changes in floral and faunal assemblages recorded in the sedimentary rocks of a region are often regarded as indications of changes in surface elevation (e.g., Xu [1981] and references cited in Powell [1986]). One problem with such inferences is that a wide variety of indicators, from oxygen isotopes in foraminifera to plant fossils from low elevations, show that there has been a pronounced cooling of Earth's surface since the beginning of the Cenozoic Era. The magnitude of this cooling depends on latitude and may have been greater than 15 °C at middle to high latitudes in the Northern Hemisphere (Shackleton and Kennett, 1975; Wolfe, 1978). A drop in surface temperature of ~10 °C appears to have occurred in the Northern Hemisphere at the end of the Eocene Epoch (Savin et al., 1975; Wolfe, 1978). Another phase of cooling appears to have begun, at least at high latitudes, in middle Miocene time (16–14 Ma) (Savin et al., 1975; Shackleton and Kennett, 1975). Wang (1984) inferred a comparable Miocene drop, of roughly 5–6 °C, in northern China. There is considerable evidence for a global cooling beginning at the end of Miocene time (Kennett, 1982) and the Pleistocene Epoch was a time of major glaciations.

A second problem associated with using terrestrial species to infer surface height changes is that some methods for estimating climatic change from the fossil record appear to be less reliable than others. Wolfe (1971) discussed critically the practice of inferring climatic conditions on purely taxonomic grounds, and advocated strongly the use of criteria based on the physiognomic characteristics of the plants in the fossil record.

Powell (1986) put forward a "model" for the tectonics of Asia that included, among other distinctive features, a 30 m.y. pause in the convergence of the Indian plate with Asia during a time (50–20 Ma) when a considerable amount of geophysical evidence suggests that it moved 1500 km northward (Molnar and Tapponnier, 1975). This model is closely linked to an assumed history of uplift of the surface of the plateau (Powell, 1986, p. 85), whose principal feature is an accelerating rise in surface height from about 1500 to 5000 m since late Miocene time, that is based on taxonomic changes in the paleobotanical record. The proposed rate of uplift is not, in itself, particularly remarkable. What makes the proposal intriguing is that, if it is correct, about 3500 m of uplift would have occurred at a time of active normal faulting and crustal thinning (Armijo et al., 1986). Isostatic compensation of a thinning crust would, in the absence of other processes, require subsidence, not uplift, of the surface. If such uplift could be demonstrated and quantified, it would place an important constraint on processes taking place in the upper mantle beneath Tibet (England and Houseman, 1989).

Unfortunately, in addition to containing mistakes of the character of those discussed above, the proposed history of uplift of Tibet is based on the following faulty inferences from paleobiological data. Many of the observations on which this history is based were made in the margins of the plateau, particularly on the north slope of the Himalaya, where the tectonic history is very different from that of the interior of the plateau; it is not necessarily correct to assume that the species preserved in deposits at the edge of the plateau inhabited its interior at the same time. Moreover, few or none of the studies took account of the simultaneous global change in climate when interpreting the local changes in flora as indicating surface uplift. For example, a drop in mean surface temperature of 10 °C owing to climatic changes since the late Miocene would, if not accounted for, yield an apparent surface uplift of about 1500 m for a lapse rate of 6 °C/km.

It is not our intention to advocate a particular history of climate or uplift of the Tibetan plateau, but to indicate that current estimates of surface uplift of the region are likely to be inaccurate, as well as fundamentally qualitative because they have not been assigned meaningful uncertainties.

CONCLUSIONS

We are aware of no reliable, quantitative estimates of rates of surface uplift in mountain ranges that place useful constraints on tectonic processes. Many observations reported as yielding rates of surface uplift in mountain ranges are in fact observations of exhumation, and hence not of uplift at all. Some observations provide reliable measures of the uplift of rocks, but because erosion rates can be high, the mean surface elevations may be decreasing while the rocks are uplifting. Finally, some inferences of uplift may be contaminated by secular climatic change.

Paleobiology appears to offer the possibility of determining quantitatively changes in the surface elevation of mountain ranges, by employing criteria based on the physiognomy of plants in the fossil record (Wolfe, 1971, 1979; Meyer, 1986), and by taking proper account of changes in climate.

Incorrect inferences of surface uplift may influence studies of dynamics of other systems than the solid Earth. It seems entirely plausible that surface uplift could profoundly affect global climate (Birchfield and Weertman, 1982). Ruddiman and Raymo (1988) suggested that a rapid increase in the elevation of mountain ranges and plateaus in the past few million years is, in fact, responsible for a fundamental change in global climate. We suggest above that the uplift to which they appeal may be more apparent than real, and the *apparent* evidence for uplift may, conversely, be the result of climatic change.

We have stressed the importance of measuring accurately rates of surface uplift because their quantification would tell us much about the mechanics of orogenic belts. If it could be demonstrated—as has been suggested several times—that large regions of the continental crust are elevated in a fashion that cannot be explained by crustal thickening alone, then we would have an important constraint on the behavior of continental lithosphere and its underlying upper mantle (e.g., England and Houseman, 1989).

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