# Paleoelevation
# GEOL5690, Tectonic History of the Western U.S.

Topography is the first-order geophysical observable. There is a lot of information there, and while easily obtained today, it is much harder to get in the geologic past provided you are above sea level. Often sedimentary deposits are rare. Exposures of bedrock don't yield a lot of information easily. This document is to provide an introduction to many of the techniques that have been employed. Most do not provide a single, absolute elevation. Virtually all these techniques require comparison between two or more sites, which means that finding valid pairs (same age, or same latitude, or similar environment) are important and potential sources of inaccuracy.

**Erosion**. The classic source of geologic inference of uplift; after all, you can't erode something until it has some elevation above base level. The flaw is that erosion might be delayed a long time, and it won't give much information about the magnitude of any surface uplift. One example was proposed by Gregory and Chase (1994), who proposed that the Southern Rockies reached modern elevations by the close of the Eocene but, under the right climatic conditions, would not see erosion until much later in the Tertiary. Erosion can be inferred from low-T geochronology, appearance of detritus downstream, and through classic geologic cross-cutting relationships. Related are arguments of tilting of sedimentary rocks (see sedimentology, below).

Gregory, K. M., and Chase, C. G., 1994, Tectonic and climatic significance of a Late Eocene low-relief, high-level geomorphic surface, Colorado: *Journal of Geophysical Research*, v. 99, no. B10, p. 20141-20160.

**Sedimentology.** Used to see if a fluvial system has been tilted (and thus the upper end raised). In a sand-bottomed and sand-sided channel, the depth of the channel and the maximum clast size are related to the gradient of the channel. Relationship used by McMillan et al. (2002) and Cassel and Graham (2011) is that of Paola and Mohrig (1996)

$$S_{est} = \frac{0.094 D_{50}}{h} \qquad (1)$$

where $S_{\text{est}}$ is the slope estimate and $D_{50}$ is the median grain size (in practice this is a median size of point samples of the deposit from a uniform grid) and $h$ is the bank-full flow depth. If the river extended normal to an axis of uplift, then the difference between modern slopes and the paleoriver would document a change in elevation (e.g., McMillan et al., 2002). If the river system extends far enough, the slope times the distance of the river would give an estimate of the elevation difference between upstream and downstream extents (e.g., Cassel and Graham, 2011)

Application of this technique requires the paleoriver to satisfy some requirements. The river must have been a braided or anastamosing stream without vegetation holding on to banks (non-cohesive banks). Application to other rivers (e.g., meandering ones) can result in large errors (e.g., Fig. 5B of Paola and Mohrig, 1996). Estimating the bank-full depth can be somewhat tricky.

Cassel, E. J., and Graham, S. A., 2011, Paleovalley morphology and fluvial system evolution of Eocene-Oligocene sediments ("auriferous gravels"), northern Sierra Nevada, California:

Implications for climate, tectonics, and topography: *Geological Society of America Bulletin*, v. 123, no. 9-10, p. 1699-1719, doi: 10.1130/B30356.1.

McMillan, M. E., Angevine, C. L., and Heller, P. L., 2002, Postdepositional tilt of the Miocene-Pliocene Ogallala Group on the western Great Plains: Evidence of late Cenozoic uplift of the Rocky Mountains: *Geology*, v. 30, no. 1, p. 63-66.

Paola, C., and Mohrig, D., 1996, Palaeohydraulics revisited: Palaeoslope estimation in coarse-grained braided rivers: *Basin Research*, v. 8, p. 243–254, doi: 10.1046/ j.1365-2117.1996.00253.x.

**Paleontology: Mean Annual Temperature.** We are all familiar with the concept of climate belts originally put forward by von Humboldt in 1807. Most often, plant paleontology is used as plants are attached to the landscape, but occasionally some arguments arise from cold-blooded vertebrates (not so much lately). So it stands to reason that identifying paleoclimate belts might provide a means of estimating elevations. This approach is reviewed by Meyer (2007). Earlier applications relied on identifying fossil plants as behaving the same as modern cousins (a "nearest living relative" model). Some, like Axelrod (1962, 1998) identified modern plant assemblages most like ancient plant fossils and then assigned the modern climatic characteristics to those assemblages. This has mostly been discarded as both climate change was ignored in many of these studies and there is little assurance that evolutionary changes would preserve climatic preferences.

A second approach is to try to exploit the characteristics of leaves that reflect climate. The most robust of those is the frequency of species of leaves without teeth, a characteristic that is well correlated with Mean Annual Temperature (MAT). Leaf shapes tied to mean annual temperature have been recognized since the early 20th century. Leaf physiognomy has been advocated as a robust estimator of climatic conditions (e.g., Meyer, 1992, Wolfe, 1993). The idea is that angiosperm leaf shape is tied to various environmental parameters (in essence, convergent evolution). For this to work, you generally need to have preservation of a number of species/genera (~30 is number I often hear). Issues include non-random preservation or collection of species, and the possibility that leaves have traveled some distance to preservation site. At one level, the community seems to have settled on mean annual temperature as a robust parameter that can be estimated from leaves. Jack Wolfe (1993) developed a more involved approach that tried to extract multiple climatic parameters from a far more thorough cataloging of leaf parameters and modern climate measurements; this produced the CLAMP multidimensional correspondence analysis database that could then be applied to fossil flora.

In general, these approaches yielded some estimate of climatic conditions that themselves do not provide a paleoelevation. Instead, these climatic conditions had to be calibrated against elevation, and this has proven troublesome. A brief review of the relation of temperature to elevation is appropriate.

First law of thermodynamics (conservation of energy) run through ideal gas law holds that as a parcel of air in adiabatic conditions changes pressure (d*P*), the total energy must remain constant:

$$c_p dT - \frac{dP}{\rho} = 0 \qquad (2)$$

Where $c_P$ is the heat capacity at a constant pressure. To get this to elevation, assume hydrostatic equilibrium in the atmosphere:

$$dP = -\rho g dz \qquad (3)$$

Which can be solved to get pressure *P* (in Pa) as a function of height *h* (in m):

$$P = 101325\left(1 - 2.25577 \cdot 10^{-5} \cdot h\right)^{5.25588} \qquad (4)$$

We combine (2) and (3) to get:

$$\begin{aligned} c_p dT &= -g dz \\ \frac{dT}{dz} &= -\frac{g}{c_p} = 9.8°\text{C/km} \end{aligned} \qquad (5)$$

Saturated or moist adiabatic lapse rates are somewhat harder to derive and can depend on the particulars of the precipitation, but basically latent heat is released as water condenses, which reduces the temperature drop by about half.

This by itself is sensible, but we aren't talking about the movement of air up and down in a weather system, we are talking about the rate of change of mean annual temperature (MAT) with elevation averaged over years. This is tricky on three scores. One is that the variation of mean annual temperature with elevation is not simply an atmospheric lapse rate; instead empirical measurements are employed. For instance, in the U.S. Rockies, MAT declines by about 7°C/km, but eastern Europe only sees a decline of about 5.5°C/km (Diaz & Bradley, 1997) and local estimates vary from ~5°C/km to ~8°C/km (Meyer, 1992). A second is the effect of being in the continent or near the ocean (figure at left), an effect nearly any near-coastal resident can attest to (going from San Francisco to Oakland to Sacramento, for instance, can entail large temperature differences, which tends to drive frequent ocean breezes that have made Pacheco Pass in the area a magnet for wind turbines). A third is that we need the equivalent values at some point in the geologic past. The result has been a confusing literature with things labeled as "lapse rates" that are not, in fact, lapse rates and linear approximations to variations that are described as

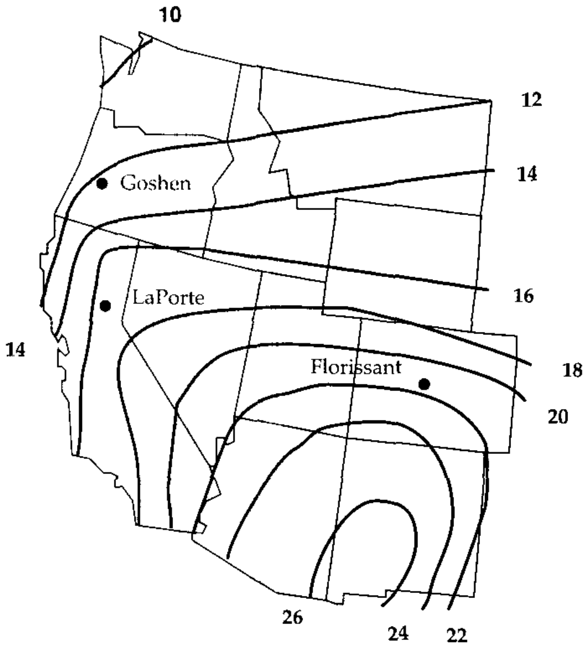

**Figure 2. Mean annual temperature (MAT) at sea level in °C for modern western United States. Note that MAT increases inland due to effects of continentality and elevated base level. Redrawn after Meyer (1986).**

"rates" that are just linear approximations. In general, application has usually been to determine some vertical rate of temperature change, apply that working from an estimate of the sea level temperature likely at the desired locale (often derived from a sea level estimate near the coast) and then making any necessary local adjustments to get an elevation.

One of the more contentious locales has been the Florissant Fossil Beds west of Colorado Springs. Originally described as a low-elevation flora by MacGinitie (1953), who collected many of the fossils, first Meyer (1992) and then Wolfe (1992) applied the physiographic approach with lapse rates to come up with paleoelevations of ~2500m. Axelrod (1998) estimating far warmer temperatures and a cooler sea level and a higher lapse rate and further inferring that the Florissant forest had to be lower than some other paleoflora came up with an estimate of only 455m. Several other workers have also had a swing at Florissant, most tending to the higher elevations.

An offshoot of these higher estimates is the need to prevent erosion from starting by the time these higher elevations would have been present, as was discussed above.

Axelrod, D. I., 1962, Post-Pliocene uplift of the Sierra Nevada, California: *Geological Society of America Bulletin*, v. 76, p. 183-198.

Axelrod, D. I., 1998, Paleoelevation estimated from Tertiary floras, *in* Ernst, W. G., and Nelson, C. A., eds., Integrated Earth and Environmental Evolution of the Southwestern United States: The Clarence A. Hall, Jr. Volume: Columbia, Maryland, Bellweather Publishing (for Geol. Soc. Amer.), p. 70-79.

Diaz, H. F., and Bradley, R. S., 1997, Temperature variations during the last century at high elevation sites: *Climatic Change*, v. 36, no. 3-4, p. 253-279, doi: 10.1023/A:1005335731187.

Meyer, H. W., 1992, Lapse rates and other variables applied to estimating paleoaltitudes from fossil floras: *Palaeogeography Palaeoclimatology Palaeoecology*, v. 99, no. 1-2, p. 71-99.

Meyer, H. W., 2007, A Review of Paleotemperature Lapse Rate Methods for Estimating Paleoelevation from Fossil Floras: *Reviews in Mineralogy and Geochemistry*, v. 66, no. 1, p. 155-171, doi: 10.2138/rmg.2007.66.6.

Wolfe, J. A., 1993, A method for obtaining climate parameters from leaf assemblages, *U.S. Geological Survey Bulletin*, v. 2040, 71 p.

**Paleontology-Moist Static Energy**. An attempt to do an end-run around the various corrections for varying lapse rates and continentality corrections was an approach advocated by Forest et al. (1999) and applied in later Wolfe papers (e.g., Wolfe et al., 1998). Moist static energy in the atmosphere is defined as the sum of moist enthalpy and GPE/unit mass, excluding kinetic energy. This is defined by Forest et al. as

$$h = c'_p T + L_V q + gZ = H + gZ \tag{6}$$

where $c_P'$ is the specific heat capacity of moist air, $T$ in K, $L_V$ is latent heat of vaporization for water, $q$ is specific humidity. $H$ is moist enthalpy. $c'_p = c_{pd}(1-q) + c_w q$, where the new heat capacities are for dry air and water. There is also a correction for the temperature dependence of latent heat of vaporization $L_v + L_{vo}(1+(T-273)(c_{pv}-c_w))$. The claim is that moist static energy is virtually conserved following air parcels (changed by radiative heating and surface fluxes of latent and sensible heat). To use this for paleoelevation, you need to be able to estimate moist enthalpy (*H*) and assume *h* is constant at a given latitude (presume zonal flow). These are calculated from a multivariate analysis of leaf shapes in the modern day (so exploiting CLAMP). Similar parameters are then used in the past. To use this, you need leaf collections along lines of atmospheric transport with one at a known (typically near sea level) elevation. Uncertanties are typically +/- 1 km. This has been applied to Florissant by Wolfe et al., (1998), yielding paleoelevations of 3.8 ± ~1 km.

Forest, C. E., Wolfe, J. A., Molnar, P., and Emanuel, K. A., 1999, Paleoaltimetry incorporating atmospheric physics and botanical estimates of paleoclimate: *Geological Society of America Bulletin*, v. 111, no. 4, p. 497-511, doi: 10.1130/0016-7606(1999)111&lt;0497:PIAPAB&gt;2.3.CO;2.

Wolfe, J. A., Forest, C. E., and Molnar, P., 1998, Paleobotanical evidence of Eocene and Oligocene paleoaltitudes in midlatitude western North America: *Geological Society of America Bulletin*, v. 110, no. 5, p. 664-678.

**Paleontology: $CO_2$ pressure**. First a little botany. Plants need carbon dioxide to live, so they need to absorb it from the atmosphere. But they also exist in an atmosphere that wants to suck out their moisture, so they have an impermeable membrane (the epidermis). Stomata are pores allowing the interaction of the plant with the atmosphere. When there is a lot of carbon dioxide, only a few stomata are needed to get the plant its $CO_2$ needs. Conversely, when there is little carbon dioxide, many more stomata are needed. The level is the partial pressure of $CO_2$, which will covary with atmospheric pressure. Although this has been used quite a bit for estimating past variations in $CO_2$ concentration in the atmosphere, it took some time for an attempt to apply to paleoelevation (McElwain, 2004; Kouwenberg et al., 2007). To apply this, you need same species (ideally) at a low, known elevation and the target uncertain elevation, though it is possible that using other estimates of sea-level $CO_2$ might be OK. It is likely that variations between modern and fossil species might present difficulties (but all this should be addressed in the paleo-$CO_2$ literature anyways). Modern observations suggesting trouble include stomatal density varying with temperature as well (in some cases overwhelming the $CO_2$ effect; Zhang et al., 2020), and with stomatal density even varying with sun aspect. While this potentially could be used on gymnosperms as well as angiosperms, it seems to have a lot of hurdles remaining.

McElwain, J. C., 2004, Climate-independent paleoaltimetry using stomatal density in fossil leaves as a proxy for CO2 partial pressure: *Geology*, v. 32, no. 12, p. 1017-4, doi: 10.1130/G20915.1.

Kouwenberg, L. L. R., Kurschner, W. M., and McElwain, J. C., 2007, Stomatal Frequency Change Over Altitudinal Gradients: Prospects for Paleoaltimetry: *Reviews in Mineralogy and Geochemistry*, v. 66, no. 1, p. 215-241, doi: 10.2138/rmg.2007.66.9.

Zhang, L., Zhang, S. R., Li, Q. J., and Quan, C., 2020, Reduced stomatal frequency with rising elevation on the Tibetan Plateau: *Global Ecology and Conservation*, v. 24, e01326, doi: 10.1016/j.gecco.2020.e01326.

**Oxygen/Hydrogen isotopes**: This is probably the most popular proxy for paleoelevation out there. Instead of temperature, the idea is to take advantage of Rayleigh distillation of water (which is open system—the water is removed as is condenses), where heavier isotopes rain out at lower elevations and so precipitation is progressively lighter the higher one goes. The theory is that at a given temperature, the partitioning of isotopes between those in precipitation, $\delta^{18}O_p$, and those in vapor $\delta\,^{18}O_v$, is

$$\delta^{18}\mathrm{O}_p = \alpha_O(T)\left(\delta^{18}\mathrm{O}_v + 1000\right) - 1000 \qquad (7)$$

recalling, of course, that

$$\delta^{18}\mathrm{O} = \left(\frac{R}{R_{\mathrm{SMOW}}} - 1\right)1000 \qquad (8)$$

where $R$ is the ratio of $^{18}O$ to $^{16}O$, and $R_{\mathrm{SMOW}}$ is the ratio in a defined standard. For oxygen, $\alpha$ increases from about 1.008 at 30°C to 1.012 near 0°C; from there to pure vapor-ice interactions values increase rapidly (~1.020 near -20°C). Somewhat similar equations exist for hydrogen and deuterium. Converting all this to values in precipitation is challenging (see review in Rowley, 2007), so often some kind of empirical calibration is used (often of river water at elevation instead of precipitation). Materials that can be used include pedogenic carbonates, carbonate cements, biological materials (e.g., teeth), lacustrine carbonates, volcanic clays, and volcanic glass rinds. Problems include having a sea level reference, knowing the isotopic composition of the source waters (and temperatures), reincorporation of evaporation/transpiration, possible seasonality to the signal, differing climates in the past (e.g., different relationships of rainout to topography), and understanding drainage basin geometry. A related approach uses biomolecules from plant parts that are interpreted for paleoelevation from the isotopic values observed (e.g., Hren et al., 2010).

Rowley, D. B., 2007, Stable Isotope-Based Paleoaltimetry: Theory and Validation: *Reviews in Mineralogy and Geochemistry*, v. 66, no. 1, p. 23-52, doi: 10.2138/rmg.2007.66.2.

Hren, M. T., Pagani, M., Erwin, D. M., and Brandon, M. T., 2010, Biomarker reconstruction of the early Eocene paleotopography and paleoclimate of the northern Sierra Nevada: *Geology*, v. 38, no. 1, p. 7-10, doi: 10.1130/G30215.1.

**Clumped isotopes (Δ47):** This is mainly a paleotemperature technique applied to carbonates, frequently pedogenic soil carbonates; the proportion of heavy isotopes of carbon and oxygen that end up bonded together in $CO_2$ is represented by $\Delta_{47}$. "Briefly, the $\Delta_{47}$ value [for $CO_2$] is the difference in per mil between the measured 47/44 ratio of the sample and the 47/44 ratio expected for that sample if its stable carbon and oxygen isotopes were randomly distributed among all isotopologues." (Ghosh et al, 2006). The proportion of bonding depends on the temperature at the time the carbonate is created, with increased clumping favored at lower temperatures. This technique also produces an improved estimate of the oxygen isotopic composition of the source waters, which can be used as above. Paleotemperatures themselves fall into the usual set of difficulties discussed for MAT paleobotany. Wrinkles here have a lot to do with when and where the carbonate was created. As these are frequently seasonal, converting temperatures to mean annual temperatures is challenging. Although seasonality is mainly discussed with respect to temperature, stream waters can have a seasonal isotopic variation as well. So far, differences in how temperatures are interpreted appear to be more of a difficulty than the actual measurements (e.g., contrasting interpretations of the Sheep Pass Frm by Lechler et al., 2013 and Snell et al., 2014).

Ghosh, P., Adkins, J., Affek, H., Balta, B., Guo, W., Schauble, E., Schrag, and D., E., J., 2006, $^{13}$C-$^{18}$O bonds in carbonate minerals: A new kind of paleothermometer: *Geochemica et Cosmochimica Acta*, v. 70, p. 1439-1456.
Lechler, A. R., Niemi, N. A., Hren, M. T., & Lohmann, K. C. (2013). Paleoelevation estimates for the northern and central proto-Basin and Range from carbonate clumped isotope thermometry. *Tectonics*, 32(3), 295–316.
Snell, K. E., Koch, P. L., Druschke, P., Foreman, B. Z., & Eiler, J. M. (2014). High elevation of the “Nevadaplano” during the Late Cretaceous. *Earth and Planetary Science Letters*, 386(C), 52–63.

**Triple Oxygen.** More recently, addition of Δ’$^{17}$O is showing up as something to help understand δ$^{18}$O. Until the past decade, the relative rarity of $^{17}$O and the absence of a clear need for it left it unstudied, but there has been a recent upsurge in using it.

$$\Delta'^{17}\mathrm{O} = \ln\left(\frac{R_{17}}{R_{17_{ref}}}\right) - 0.528\ln\left(\frac{R_{18}}{R_{18_{ref}}}\right) \tag{9}$$

Where $R_{17}$ is the observed ratio of $^{17}$O to $^{16}$O and the *ref* subscript is a reference value; these Δ’$^{17}$O values are often reported as per meg (i.e., actual value x 10$^{6}$), showing just how small these variations are. The reference value largely reflects the effects of Rayleigh distillation on waters so that Rayleigh distillation, often significant for δ$^{18}$O as described above, has no signal in Δ’$^{17}$O. Instead, evaporation has a profound effect on Δ’$^{17}$O, making it very negative, which is the main application in paleoaltimetry. The other main effect is that seawater is mildly negative while freshwater is positive, allowing for recognizing these two possible sources. The δ$^{18}$O values extracted from clumped isotopic work reflect the composition of the water from which the carbonate was precipitated, but that water need not have the same ratio as the freshwater in the area. By employing Δ’$^{17}$O, the δ$^{18}$O values can then be corrected for evaporation. This may still leave other ambiguities (e.g., was it river water from on high or local precipitation?). One example explored the Goler Formation in southern California (that was quite troublesome in the Lechler et al. paper cited above). In this case, the isotopic composition of water creating marine shells was, when corrected, negative enough to suggest freshwater derived from highlands in the drainage basin (Kelson et al., 2022).

Kelson, J. R., Petersen, S. V., Niemi, N. A., Passey, B. H., and Curley, A. N., 2022, Looking upstream with clumped and triple oxygen isotopes of estuarine oyster shells in the early Eocene of California, USA: Geology, v. 50, no. 7, p. 755-759, doi: 10.1130/G49634.1.

**Volcanic vesicles**: Rather like the measurement of stomata in leaves, the idea is to recover the paleoatmospheric pressure and thus the elevation. The trick is to compare the sizes of vesicles at the top and bottom of a flow and then noting that the ratio in the volumes of top ($V_t$) to bottom

($V_b$) equals to the ratio of the atmospheric pressure $P$ + the weight of the flow to the pressure alone, i.e.

$$\frac{V_t}{V_b} = \frac{P + \rho g H}{P} \tag{10}$$

This has nearly always been applied to basalts, and smooth surfaced ones at that (pahoehoe flows) (Sahagian et al., 2002a). Volumes are measured with X-ray tomography in newer studies. Obviously only applicable to places with the right source materials. Known complications are if the flow thickened or thinned during emplacement (e.g., Bondre, 2003), and it is evidently assumed that the distribution of masses within the bubbles is identical top and bottom. While this has been applied to much of the periphery of the Colorado Plateau (Sahagian et al., 2002b).

Sahagian, D. L., Proussevitch, A. A., and Carlson, W. D., 2002a, Analysis of vesicular basalts and lava emplacement processes for application as a paleobarometer/paleoaltimeter: *The Journal of Geology*, v. 110, no. 6, p. 671-685, doi: 10.1086/342627.
  Bondre, N. R., 2003, Analysis of vesicular basalts and lava emplacement processes for application as a paleobarometer/paleoaltimeter: A discussion: *The Journal of Geology*, v. 111, no. 4, p. 499-502, doi: 10.1086/375279.
  Sahagian, D. L., Proussevitch, A. A., and Carlson, W. D., 2003a, Analysis of vesicular basalts and lava emplacement processes for application as a paleobarometer/paleoaltimeter: Reply: *The Journal of Geology*, v. 111, no. 4, p. 502-504, doi: 10.1086/375278.

Sahagian, D., Proussevitch, A., and Carlson, W., 2002b, Timing of Colorado Plateau uplift: Initial constraints from vesicular basalt-derived paleoelevations: *Geology*, v. 30, no. 9, p. 807-810.
  Libarkin, J. C., and Chase, C. G., 2003, Timing of Colorado Plateau uplift: Initial constraints from vesicular basalt-derived paleoelevations: Comment and Reply: COMMENT: *Geology*, v. 31, no. 2, p. 191-191, doi: 10.1130/0091-7613(2003)031<0191:Tocpui>2.0.Co;2.
  Sahagian, D., Proussevitch, A., and Carlson, W., 2003b, Timing of Colorado Plateau uplift: initial constraints from vesicular basalt-derived paleoelevations: Reply: *Geology*, v. 31, no. 2, p. 192-192.

**Cosmogenic isotopes**. Basically, cosmic rays penetrate less and less deeply into the atmosphere the closer you get to sea level. Thus the production of cosmogenic isotopes depends on elevation. Very difficult to use owing to the rapid decay of such isotopes and the various uncertainties of their production (variations in the Earth's field, strong latitudinal and elevation gradients in production, need to constrain erosion of surface). This has been used to argue for long term elevation of the Transantarctic Mtns (Riihimaki and Libarkin, 2007), but the requirements for its use are rather stringent.

Riihimaki, C. A., and Libarkin, J. C., 2007, Terrestrial cosmogenic nuclides as paleoaltimetric proxies: *in* Paleoaltimetry: Geochemical and Thermodynamic Approaches, M. J. Kohn ed., *Reviews in Minerology and Geochemistry*, v. 66, p. 269-278, doi: 10.2138/rmg.2007.66.11.

**Low-T geothermometry**. While low-temperature geothermometry is wildly popular at the moment, its applications to paleoaltimetry are more subtle. In essence, variations in ages reflecting variations in relief are then exploited for estimating regional elevations. The best example is from House et al. (1998). A transect parallel to the crest of the Sierra Nevada was sampled at near-constant elevation. Cooling dates from (U-Th)/He on apatite vary quite a bit (extremes from 40-80 Ma, most 50-75 Ma), suggesting some sites cooled below the closure temperature far earlier than other sites. From this data, the authors inferred local relief of river canyons of 2-4 km, which suggests upland elevations of at least 2-4 km and a crest elevation over 4 km as far back as 70 Ma. (Some of the assumptions here were attacked by both Clark et al. 2005 and Jones et al. 2004).

A similarly inventive application of low-T geochronology was explored by McPhillips and Brandon (2012). They used all available low-T geochronology as well as Al-in-Hornblende geobarometry to see how relief and elevation have changed over the Tertiary in the Sierra Nevada. They find an early high Sierra that then decayed until rejuvenated starting ~20 Ma.

Clark, M. K., Maheo, G., Saleeby, J., and Farley, K. A., 2005, The non-equilibrium landscape of the southern Sierra Nevada, California: *GSA Today*, v. 15, no. 9, p. 4-10, doi: 10:1130/1052-5173(2005)015<4:TNELOT>2.0.CO;2.

House, M. A., Wernicke, B. P., and Farley, K. A., 1998, Dating topography of the Sierra Nevada, California, using apatite (U-Th)/He ages: *Nature*, v. 396, no. 6706, p. 66-69.

Jones, C. H., Farmer, G. L., and Unruh, J. R., 2004, Tectonics of Pliocene removal of lithosphere of the Sierra Nevada, California: *Geological Society of America Bulletin*, v. 116, no. 11-12, p. 1408-1422, doi: 10.1130/B25397.1.

McPhillips, D., and Brandon, M. T., 2012, Topographic evolution of the Sierra Nevada measured directly by inversion of low-temperature thermochronology: *American Journal of Science*, v. 312, no. 2, p. 90-116, doi: 10.2475/02.2012.02.