
Palaeoaltimetry from energy conservation principles

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A KNOWLEDGE of past changes in the mean elevations of large continental areas is important for understanding both dynamic processes in the Earth's mantle¹ and the evolution of the global climate². But virtually all methods for determining palaeoelevations are problematic³, in part because changes in either elevation or climate can give rise to the same observed geological phenomena⁴. Ideally, palaeoelevations would be inferred directly from estimates of palaeopressure, and it has recently been shown⁵ that vesicles in basaltic lava flows preserve a record of atmospheric pressure at the altitude of emplacement; however, the elevations thus obtained have large errors (~1.4 km), and the method can at present be applied only to lavas that have had a simple emplacement history⁵. Palaeobotanical methods for estimating palaeoelevation have received much attention^{6–12}, but they rely on empirical temperature–elevation relationships, the uncertainties in which are difficult to evaluate for past climates. Here we describe an alternative palaeobotanical approach, based on energy conservation in the atmosphere, in which fossil leaf assemblages are used to infer enthalpy, rather than temperature. This approach is relatively insensitive to palaeoclimate, with an expected error in palaeoelevation of ~700 m.

The use of fossil leaf assemblages for palaeoaltimetry consists of two steps: first, estimating from fossil plants a climate parameter that varies with altitude (like mean annual temperature), and second, using differences in that parameter from separate sites to estimate elevation differences⁶, given *a priori* knowledge of the variation of the parameter with altitude. Mean annual temperature has been used predominantly as the climate param-

eter because it shows correlations with physiognomic characteristics of leaf samples from living plants^{13,16}. Assuming that characteristics of fossil leaf assemblages obey the same relationship, palaeotemperature estimates can be inferred^{7,17} with uncertainties perhaps as low as 1 °C. As mean annual temperature varies with altitude, latitude and geological time, palaeoelevation (Z) may be calculated by comparing surface temperatures, T , of high- and low-altitude locations at the same latitude and of the same age using:

$$Z = (T_{\text{high}} - T_{\text{low}}) \gamma^{-1} \quad (1)$$

where $\gamma = -\partial T / \partial Z$ is the empirical relationship between mean annual temperature at the surface and altitude derived from the present climate. (Note that γ is not a gradient of T within the atmosphere and so is not a lapse rate in the meteorological sense, although it has been termed a 'terrestrial lapse rate'.)

Not only do relationships between T and Z lack a firm theoretical foundation, but large spatial variations also exist in present-day estimates of γ in the western United States⁸⁻¹⁰. Meyer¹⁰ calculated local estimates of γ for 39 areas of the world with surface topography >750 m and found $\gamma = 5.9 \pm 1.1 \text{ K km}^{-1}$, but with a range of 3.64–8.11 K km^{-1} . Moreover, we have no reason to expect that the present laterally varying empirical relationships should hold in the different climates that occurred in geological time. We therefore seek a measurable thermodynamic quantity whose distribution with altitude and longitude is constrained well by both theory and observations. The use of a quantity chosen to agree with fundamental thermodynamic laws is obviously more reliable than one fitted empirically to data spanning a fraction of the twentieth century—a small fraction of geological history.

Derived from the first law of thermodynamics, moist static energy^{18,19}, h , is the total specific energy content of air ($\sim 300 \text{ kJ kg}^{-1}$ in mid-latitudes), excluding kinetic energy, which is very small (<5%) compared with the other terms:

$$h = c_p' T + L_v q + gZ = H + gZ \quad (2)$$

where c_p' is the specific heat capacity at constant pressure of moist air, T is temperature (in K), L_v is the latent heat of vaporization for water, q is specific humidity, g is the gravitational acceleration, Z is altitude and H is moist enthalpy. The specific heat capacity of moist air, $c_p' = (1-q)c_{pd} + qc_{pv}$, accounts for compositional changes of the air, where c_{pd} and c_{pv} are the speci-

FIG. 2 The spatial distribution of mean annual moist static energy. Crosses indicate station locations from the ISMCS dataset²². Mean moist static energy h was calculated by independently calculating a mean temperature and mean specific humidity and by combining the respective sensible heat, latent heat and potential energies. We accounted for the temperature dependence of the latent heat of vaporization, $L_v = L_{v0} + (T - 273.15)(c_{pv} - c_l)$, where c_{pv} and c_l are the specific heat capacity at constant pressure of water vapour and liquid water, respectively, and L_{v0} is the latent heat of vaporization at 0 °C. This second-order effect and the dependence of c_p' on q are most important in warm and moist climates, where they contribute variations of $\sim 5\text{--}10 \text{ kJ kg}^{-1}$ corresponding to apparent differences in elevation of 500–1,000 m. Of the various averaging periods, we chose mean annual h based on the small error from zonal asymmetries and on the ability to predict mean annual enthalpy from fossil leaf physiognomy. The locations of the fossil collection sites are indicated by filled circles: Goshen, Oregon; LaPorte, California; and Florissant, Colorado. Inset, the variation of h with latitude (in degrees). Limiting the data to 30°–60° N yields the estimate of $s_h = 4.5 \text{ kJ kg}^{-1}$.

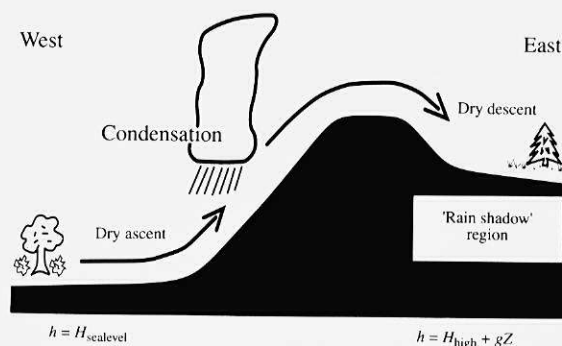
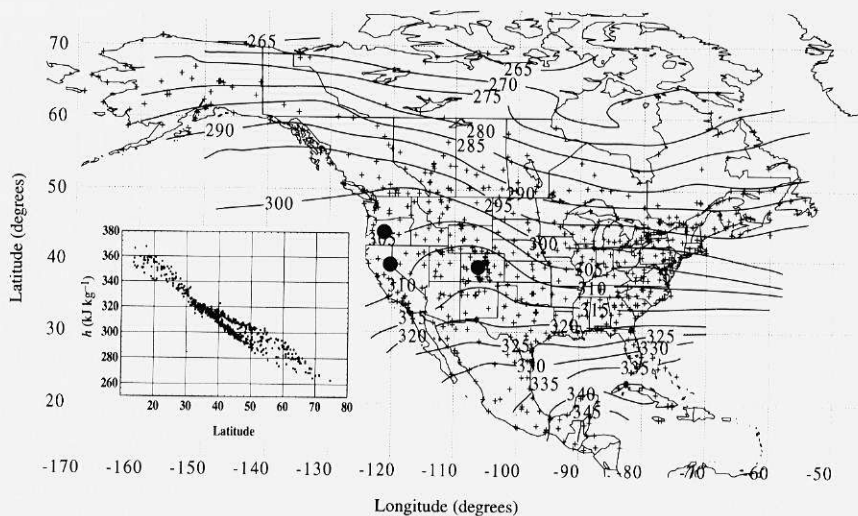


FIG. 1 Surface air conserves moist static energy as it traverses a mountainous region through conversions between sensible heat, latent heat and potential energy. During dry ascent and descent, changes in sensible heat compensate for changes in potential energy. For condensation processes at a constant height, the sensible heat increases when the latent heat of water vapour is released. The potential energy for the high-elevation site is the difference in moist enthalpy, $\Delta H = H_{\text{seal level}} - H_{\text{high}}$, which, via $\Delta H/g$, yields the elevation estimate.

fic heat capacities of dry air and water vapour respectively. Although the internal energy, $c_p' T$, is the largest term in equation (2), typical variations in the three separate components are roughly of equal magnitudes, $\sim 30\text{--}50 \text{ kJ kg}^{-1}$, for the ranges of T , q and Z of approximately 0–30 °C, 0–20 g kg^{-1} , and 0–4,000 m respectively.

Two properties of moist static energy make it a desirable candidate for inferring palaeoelevations. First, changed only by radiative heating and surface fluxes of latent and sensible heat, it is nearly conserved following air parcels and hence is approximately constant along trajectories. Second, the value of h in the atmospheric boundary layer (the lowest 1–1.5 km) is usually constrained by convection to be close to the value of h in the upper troposphere^{20,21}. (The actual constraint is along vortex lines rather than in the vertical direction, but the difference is usually small.) The Earth's rotation causes large-scale movement in the upper troposphere to flow nearly from west to east at mid-latitudes, so that contours of h there should also trend nearly west–east. This implies that h at the surface should be nearly invariant with longitude (that is, be zonally symmetric).

As a caveat, the large-scale atmospheric flow is influenced by topography that will affect the zonal symmetry. So when applying this technique, one should ascertain the zonal symmetry of the palaeoclimate; this requires a better understanding of stationary waves in the general circulation of the atmosphere.

Assuming that h is zonally symmetric along the Earth's surface, if we can estimate enthalpy at sea level ($H_{\text{sea level}}$) for a particular latitude, it follows from equation (2) that the palaeoaltitude, Z , of another location at the same latitude is given by

$$Z = \frac{H_{\text{sea level}} - H_{\text{high}}}{g} \quad (3)$$

where H_{high} is the enthalpy at the high-altitude locations (Fig. 1). Therefore, the necessary assumptions to estimate palaeoaltitude are: (1) the surface moist static energy is zonally symmetric, and (2) the estimate of enthalpy for the regional palaeoclimate is accurate.

To examine deviations from zonal invariance of h , we examined the distribution of mean h across the North American continent (Fig. 2). Using climate data for 699 surface weather stations²², we calculated mean h by independently calculating a mean temperature and mean specific humidity and by combining the respective sensible heat, latent heat and potential energies. We used averaging periods ranging from a season to a year, and various formulations of mean temperature and mean specific humidity. We quantified the longitudinal invariance by statistically fitting h to monotonic arbitrary functions of latitude, ϕ . Because we are testing for invariance, the standard deviation of the fit is more important than the actual latitude dependence. To minimize the expected error, we limited the data to 30°–60° N and divided them into three latitude bins of 10°. We linearly fitted the data within each band and used deviations from each to obtain a mean square error (Fig. 2). This approach reduces the expected error from zonal asymmetries of h , s_h , to 4.5 kJ kg⁻¹. Dividing s_h by g yields an estimate of the minimum error in altitude of 460 m.

We can compare this error to that deduced assuming the invariance of the terrestrial lapse rate, and therefore the longitudinal invariance of $G = T + \gamma Z$. For $\gamma = 5.9 \text{ K km}^{-1}$ (ref. 10), we calculated deviations from longitudinal invariance of G to be 3.2 K for the present-day climate of North America. Hence a minimum

standard error for estimating altitude can be computed from these results, from $s_z^2 = s_G^2/\gamma^2 + ((\Delta T)^2 s_\gamma^2)/\gamma^4$ where s_G and s_γ are the standard errors of G and γ respectively. With $s_\gamma = 1.1 \text{ K km}^{-1}$ (ref. 10), minimum uncertainties in the estimated altitudes increase with altitude: $s_z = 540 \text{ m}$, 660 m and 920 m for altitudes of 0 m, 2,000 m and 4,000 m, respectively. So even ignoring the unpredictable temporal changes in γ over geological time, these estimates of minimum uncertainties exceed that of 460 m from assuming longitudinal invariance in h .

The task remains to determine mean annual enthalpy from plant physiognomy. We present results of an analysis relating foliar physiognomic characters to mean annual enthalpy that exploits the method and data developed by Wolfe⁷. We searched among an array of 29 foliar characteristic scores (p_i ; see below) from 92 sites in North America for linear combinations of the foliar characteristics that covary with the local climates. At a given site, the scores are calculated as a percentage of species exhibiting each of 29 foliar characteristics: seven ranges of size, five length-to-width ratios, overall shape (obovate, elliptic, ovate, lobed), basal shape (cordate, round, acute), apical shape (emarginate, rounded, acute, attenuate), and teeth (no teeth, regular, close, round, acute, compound). We determined which foliar characteristics covary with one another (using principal components analysis) and which best correlate with enthalpy (using principal components analysis with regression²³). This creates a predictive equation of the form:

$$H = H_0 + \sum_i \alpha_i p_i \quad (4)$$

where α_i is the regression coefficient of the i th foliar characteristic. The standard error for predicting H is 3.9 kJ kg⁻¹ (Fig. 3), corresponding to a minimum error in palaeoelevation of 400 m.

To illustrate the use of this method in determining palaeoelevation, we estimated H from the physiognomic characteristics for three well known fossil floral assemblages: Florissant (34.1 Myr) in central Colorado, Goshen (32 Myr) from central Oregon and LaPorte (33 Myr) from northern California (Fig. 2). Each assemblage of fossil leaves has been appropriately scored using the same characteristics as in Wolfe's data set⁷ and has been used in previous estimates of Florissant's palaeoelevation including 2.3–3.3 km (refs 11, 12), 2.5 km (ref. 8) and 2.9 km (ref. 9). Using Wolfe's scores (personal communication) for the Goshen and LaPorte floras and Gregory's (ref. 11) for the Florissant flora, we calculated mean enthalpies of 294.6, 321.1 and 321.2 kJ kg⁻¹, respectively, for these assemblages. Assuming that both Goshen and LaPorte were deposited at sea level yields a sea-level value of mean enthalpy of 321.1 ± 2.8 kJ kg⁻¹. The difference in enthalpy between sea level and Florissant yields a palaeoelevation of 2,700 ± 670 m. Sea level 32–34 Myr ago was 180 m higher²⁴ than at present, so the palaeoelevation is 2.9 ± 0.7 km with respect to the present geoid. This estimate of Florissant's palaeoaltitude agrees with other recent estimates of palaeoaltitude, and therefore supports the contention of Gregory and Chase^{11,12}, Meyer⁸ and Wolfe⁹, that Florissant was high in Late Eocene/Early Oligocene time. As these workers have noted, there is a widespread belief that the Rocky Mountains were at a low elevation throughout much of the Late Cenozoic and rose early in Quaternary times. More importantly here, the 700-m uncertainty is smaller than those of other palaeoelevation estimates. □

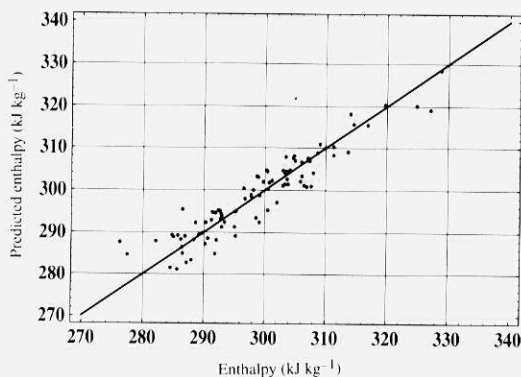


FIG. 3 The predicted mean enthalpy versus observed mean enthalpy for the annual mean, as given by equation (4) for the 92 vegetation sites (see text). The regression analysis accounts for 88% of the variations in moist enthalpy, yielding a standard error $s_H = 3.9 \text{ kJ kg}^{-1}$. Because humidity data are not available for each vegetation site, mean annual enthalpy for the plant collection sites is interpolated from nearby stations using the standard distance-weighted mean, $h_{\text{plant site}} = (\sum w_i h_i) / (\sum w_i)$ where $w_i = \exp(-r_i^2/d^2)$, where r_i is the radial distance from plant site to the meteorological station i , and d is the average distance to the nearest station, 90.8 km. Knowledge of the elevation, Z , and mean annual temperature, T , at the vegetation sites allows the calculation of mean moist enthalpy, $H = h - gZ$, from the interpolated value of h .

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