Unroofing Maine: Relating pressure of crystallization, thermochronological data, tectonics, and topography

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1. Introduction

When we admire the rounded hills and low mountains characteristic of much of Maine today, we rarely think of the rugged peaks that once were here or of the immense thickness of rock that has been eroded away to produce the present topography. To explore the character of this ancient topography and its evolution, we collected published data on cooling ages and equilibration pressures of plutons and metamorphic assemblages, now exposed at the surface, from 64 sites in Maine, New Hampshire, and New Brunswick. Then, using an average crustal thermochronological data set (or CBT), provide temperatures, at several times in the past, of rocks presently exposed at the surface. Using our Paleozoic geothermal gradient, we transformed these data into a time series of mean topographic elevations. This led us to insights into likely series of erosion.

We also used the pressure data to estimate the Paleozoic geothermal gradient in the Casco Bay area of Maine, where West et al. (1993) and West and Roden-Tice (2003) obtained mineral and apatite fission track ages. Their data, hereinafter referred to as the Casco Bay thermochronological data set (or CB), provide temperatures, at several times in the past, of rocks presently exposed at the surface. Using our Paleozoic geothermal gradient, we transformed these data into a time series of mean topographic elevations. This led us to insights into likely topographic and tectonic history.

We begin by summarizing relevant aspects of Maine’s geologic history.

2. Geology of Maine

2.1. Formation of peri-Gondwanan terranes

In the early Paleozoic the core of present-day Maine lay somewhat south of the equator on the margin of Laurentia. The coast faced...
southward toward an ocean, on the opposite side of which lay Gondwana (e.g., Hatcher, 2010, Fig. 4). The ocean was complex. Within it were arc systems as well as several microcontinents rifted from Gondwana. Collectively, these are referred to as peri-Gondwanan terranes (Hatcher, 2010). Throughout the early to mid-Paleozoic, in a series of transpressional collisions, these exotic terranes converged with and were accreted to the Laurentian margin (e.g., Hogan and Sinha, 1989, p. 24; van Staal et al., 2009; Hatcher, 2010; Hibbard et al., 2010).

2.2. Acadian orogeny

In Maine, the first of these collisions resulted in the Ordovician Taconic orogeny, recorded in rocks in the northwestern part of the state. Our earliest geobarometric data, however, are from the Late Silurian and Devonian (420–380 Ma) Acadian orogeny. The Acadian involved tectonism and broad-scale regional metamorphism, most intense in western Maine and diminishing in intensity to the east and northeast (De Yoreo et al., 1989, p. 172).

Recognizing that some details of the Acadian are debated, particularly the nature of the subduction between the last colliding microcontinent and Laurentia (Bird and Dewey, 1970, Fig. 9B; Osberg, 1978; Bradley, 1983; Eusden et al., 2000; van Staal et al., 2009; Hibbard et al., 2010), Bradley (1983, p. 391, Fig. 6) and Bradley et al. (2000) suggested the following sequence of events: (i) At ~420 Ma, deformation peaked along the present Maine coast. Simultaneously, structures along the northwestern edge of the Merrimack trough (Fig. 1) and sporadic volcanism in the Piscataquis volcanic arc indicate the beginning of northwest-dipping subduction beneath Laurentia. (ii) Somewhat later, structures along the southeastern margin of the Merrimack trough suggest the initiation of southeastward subduction (Fig. 2). Magma rising above this down-plunging plate resulted in syntectonic plutons, dating from 420 to 400 Ma, in (present-day) coastal Cambro-Ordovician rocks. (iii) The deformation front migrated northwestward across Maine over the next 40 m.y. (Fig. 1, thin lines with ages in italics), affecting sediments in the Merrimack trough and resulting in extensive plutonism in the Piscataquis volcanic arc between 407 and 397 Ma. (4) The front finally died out in southeastern Quebec at ~380 Ma (Bradley and Tucker, 2002, p. 484, Fig. 2).

Along the (present-day) coast, additional plutonism occurred between 370 and 390 Ma, thus postdating the Acadian deformation peak there (Fig. 2) (Holdaway et al., 1982, 1988; Hogan and Sinha, 1989, p. 22; Holdaway, 2004). Hibbard et al. (2010) referred to this as the Famennian event and suspected that it is related to dextral transpression and docking of a later terrane, the Meguma terrane. Contact metamorphism is associated with these plutons. Many of our pressure data come from this period (Fig. 3).

2.3. Norumbega fault zone

During the Late Paleozoic, continued oblique convergence of previously accreted terranes produced a series of NE-trending, dextral transcurrent faults and shear zones that, in Maine, have been collectively referred to as the Norumbega fault zone (Ludman and West, 1999). The Norumbega is a major transpressional structural feature, 25 to 40 km wide and extending at least 400 km from central New Brunswick to southwestern Maine. It may be connected with other faults reaching as far north as St. Lawrence Bay and as far south as southeastern Connecticut (Ludman and West, 1999). Although roughly coincident with the suture between Laurentia and Avalonia, its role as a major terrane boundary is unclear (Ludman and West, 1999).

Little evidence has been found for activity on the Norumbega before ~384 Ma, the age of the Deblois pluton in eastern Maine that is offset by it (Ludman et al., 1999). Most of the possibly 100–150 km (Ludman et al., 1999; Swanson, 1999) of dextral displacement on it appears to have occurred between ~384 and ~360 Ma (Ludman et al., 1999; West, 1999), although recent analyses suggest that some localized dextral shearing continued up to the early Mesozoic (M. Swanson, University of Southern Maine, written communication, 2012). West et al. (1993) and West and Roden-Tice (2003) have argued for significant Mesozoic dip-slip reactivation on it in the Casco Bay region. We expand upon this hypothesis below.

2.4. Alleghanian orogeny

At ~320 Ma, Gondwana began to encounter the amalgamated peri-Gondwanan/Laurentian terrane of Maine and Nova Scotia, resulting in the Alleghanian orogeny (Lux and Guidotti, 1985; Culshaw and Liesa, 1997; Hatcher, 2002). Contact first occurred in the north and then, as Gondwana rotated clockwise relative to Laurentia, progressed southward, resulting in dextral shear throughout the Appalachians. The merged continents formed the supercontinent, Pangaea. The Alleghanian appears to have been little more than a gentle bump in Nova Scotia (Murphy and Collins, 2008), but its intensity increases southward; it may have been responsible for thrust nappes in Connecticut (Wintsch et al., 2003), and it resulted in extensive folding and overthrust faulting in the central and southern Appalachians.

In Maine, the Alleghanian is represented by a number of plutons and associated deep-seated metamorphism in the southwestern part of the

Fig. 1. Map of Maine showing locations of mineral assemblages used to estimate pressures (dots). Also shown are times of crystallization (contours labeled 330 and 390) and of passage of the Acadian deformation front (thin solid lines, italic ages), the location of West's Casco Bay site, and the approximate location of the seismic line discussed by Unger et al. (1987) and Stewart (1989). M = Miramichi, MW = Mt. Washington. Deformation front data are from Bradley et al. (2000, Fig. 10). Terrane subdivisions are from Bradley (1983, Fig. 2).

Map of Maine showing locations of mineral assemblages used to estimate pressures (dots). Also shown are times of crystallization (contours labeled 330 and 390) and of passage of the Acadian deformation front (thin solid lines, italic ages), the location of West's Casco Bay site, and the approximate location of the seismic line discussed by Unger et al. (1987) and Stewart (1989). M = Miramichi, MW = Mt. Washington. Deformation front data are from Bradley et al. (2000, Fig. 10). Terrane subdivisions are from Bradley (1983, Fig. 2).
state and, quite likely, by crustal thickening there. Several of our younger geobarometric sites are of this age (Fig. 3).

2.5. Mesozoic rifting

During the Late Triassic (220 Ma) NW–SE extension began to open grabens in a linear NE–SW (present coordinates) belt, hundreds of kilometers long, in the middle of this Pangean supercontinent (Rankin, 1994, p. 177). Several of these grabens, now filled with Upper Triassic and Lower Jurassic (220–190 Ma) continental sediments, lie beneath the present Gulf of Maine (Ballard and Uchupi, 1972) and elsewhere along the eastern margin of the Appalachians (McHone and Butler, 1984; Stewart, 1989; Rankin, 1994). Along the axis of this belt, oceanic crust began to form at ~190 Ma (Vogt and Einwich, 1979; Vogt and Tucholke, 1979; LeRoy and Piqué, 2001, pp. 377–378), or possibly as late as 165 Ma (Sclater et al., 1977). However, the bordering continental crust northwest of the axis remained above sea level throughout most of the Mesozoic. Episodic submergence began in the Late Cretaceous and became more persistent in the Eocene (Uchupi and Bolmer, 2008).

The spreading was accompanied by crustal thinning. Seismic studies indicate that the crust is presently ~34 km thick southeast of the Norumbega fault zone (Fig. 1). Over a distance of ~15 km, starting ~5 km northwest of the fault zone, it gradually thickens to ~38 km. A similar gradual thickening to ~41 km takes place ~60 km farther northwest, near the northwestern edge of the Merrimack trough (Unger et al., 1987; Stewart, 1989).

3. Data and methods

3.1. Geobarometric data

We found published data on the age and equilibration pressure of mineral assemblages from 64 sites in Maine and adjacent areas (Appendices A, B). The rocks range in age from 293 to 420 Ma (Fig. 3). These assemblages largely postdate the Acadian deformation and several coincide with the Alleghanian. We assumed that the pressures are those that existed at the ages given and that the pressures reflect depths of burial, in accord with Airy isostasy. In some cases, retrograde metamorphism during initial unroofing may have obliterated evidence for earlier higher pressures.

Estimates of pressures of crystallization have uncertainties of ±0.3 kb at best; most are higher. Early ones, in particular, are ±1 kb or more. The uncertainty in ages is small in comparison, averaging ~±12 m.y., or ~±3% of the time since the Acadian orogeny.

3.2. The Casco Bay thermochronological (CBT) data set

West et al. (1993) analyzed 41 mineral separates from rocks on either side of the Norumbega fault zone near Casco Bay, southwestern Maine, using 40Ar/39Ar incremental heating. Such work yields the ages of certain minerals at the times when the temperature fell through the closure temperature for that particular mineral. The minerals analyzed were hornblende (480 ± 20 °C), muscovite (320 ± 20 °C), biotite (280 ± 20 °C), and K-spar [225 ± 25 °C (max) and 150 ± 25 °C (min)]. Ages ranged from 360 to 190 Ma. Later, West and Roden-Tice (2003) collected seven additional samples from the same area and extracted apatite fission track thermal histories from two of them. The age is the time when the rock cooled below ~90 °C and is based on the density of fission tracks. The thermal history is derived from the length distribution of tracks by Monte-Carlo simulations and is a likely, but not a unique, solution.

The Casco Bay site is located near a ‘restraining bend’ in the Norumbega fault zone (Swanson, 1999). The restraining bend extends from slightly west of Penobscot Bay to somewhere west of the Casco Bay site (Fig. 1). During dextral movement on the fault, this bend resulted in rocks on the northwest side being mushed northwestward, forming a flower structure (Fig. 4) (Swanson, 1999, p. 91). This structure appears to have begun to develop at least as early as middle Devonian and to have been particularly active throughout the Pennsylvanian and into the early Permian, up to 290 Ma (Swanson, 1999, pp. 90, 99–100). We infer that rocks on both sides of the fault zone were thickened in this process but that the thickening was greater on the northwest side. As Swanson (1999, p. 241) noted, and as we discuss below, this is consistent with the CBT data showing that rocks currently

Fig. 2. Conceptual model of collision between Avalonia and Laurentia in Late Silurian and Early Devonian time. After Bradley (1983, Fig. 6) Bradley et al. (2000, Fig. 3), Eusden et al. (2000, Fig. 4c), and Tucker et al. (2001, Fig. 11).

Fig. 3. Histogram of ages of crystallization of mineral assemblages yielding geobarometric data. E, M, L = early, middle, late.
exposed at the surface on the northwest side were consistently hotter than those on the southeast side throughout the Paleozoic.

3.3. Relating topography to depth of burial

At each geobarometric data site, let $H(t)$ be the elevation above modern sea level, averaged over an area of order 10$^2$ km$^2$, at time $t$. If $H_p$ is the known height at some time $t_p$, and $\Delta H$ is the additional height that would result from adding to this topography the rock eroded away between time $t$ and time $t_p$, then,

$$H(t) = H_p + \Delta H$$

For $H_p$, we use the known present elevation of the land surface, so $t_p = 0$ BP and ancient land surface elevations are implicitly referenced to current sea level.

To estimate $\Delta H$ we balance forces at the bases of two columns (Fig. 5), one representing the situation at some time in the past with $h_b$ meters of rock above the relevant mineral assemblage and one representing conditions today with that assemblage exposed at the surface. This yields $\rho_c(h_b + \Delta h) = \rho_ch_1 + \rho_mh_m$ where $h_1$ and $h_m$ are defined in Fig. 5, $\rho$ is density, and the subscripts $c$ and $m$ refer to crust and mantle, respectively. Because $h_b + h_1 = \Delta H + h_1 + h_m$, this simplifies to

$$\Delta H = \left(1 - \frac{\rho_c}{\rho_m}\right)h_b$$

(2)

With $\rho_m = 3300$ kg m$^{-3}$ and $\rho_c = 2750$ kg m$^{-3}$, $\Delta H = h_b/6$. Uncertainties in pressures of crystallization result in uncertainties in $\Delta H$ of $\pm 185$ to $\pm 620$ m. Pressure estimates at neighboring sites, however, were generally comparable, giving us some confidence in our paleotopographic reconstructions below.

3.4. Erosion laws

To recreate the land surface at any specific time in the past, we need to normalize elevations to that date. This requires a relation for the change in $H$ with time, such as

$$\frac{dH}{dt} = -E_o - E(t) + U + c$$

(3)

where $E_o$ is the chemical denudation rate, $E(t)$ is the erosion rate that is expected to decrease with time as $H$ decreases, $U$ is the regional uplift rate resulting from isostatic adjustment, and $c$ includes tectonic adjustments such as crustal thickening or slip on faults. From Fig. 5 and using Eq. (2):

$$U = \frac{d}{dt}(h_b - \Delta H) = -\frac{\rho_c}{\rho_m}\frac{dh_b}{dt}$$

(4)

The minus sign reflects the fact that erosion results in isostatic rock uplift.

Combining Eqs. (1) and (2), differentiating, combining the result with Eqs. (3) and (4), and using the density values given above yields

$$\frac{dH}{dt} = -\frac{1}{6}(E_o + E - c)$$

(5)

$E_o$ is normally small and $c$ is commonly episodic, so initially we neglect both. Cases with $c \neq 0$ are discussed later. Suffice it to say here that, without this simplification, analytical solutions become unmanageable and in one case, apparently impossible to obtain. Eq. (5) thus becomes

$$\frac{dH}{dt} = \frac{E}{6}$$

(6)

To integrate Eq. (6), we need an erosion law relating $E$ to $H$. Of course $H$ is only one of the variables affecting erosion rate. Among others are local climate, lithology, and vegetation (to the extent that it is independent of climate). In our case, at 400 Ma the present east coast of North America was oriented east–west and was well south of the equator (Hatcher, 2010). During Maine’s subsequent travels northward, it passed through many climate zones as well as major changes in Earth’s overall climate and vegetation. In equatorial regions, precipitation and weathering rates would have been higher, while in the horse latitude deserts both would have been lower. The mountains were subjected to winds from different directions that, owing to orographic effects, would have affected precipitation patterns. Finally, the many kilometers of rocks eroded included sedimentary facies, both weak and resistant metamorphic rocks, and plutonics. Because $H$ is the only one of the

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**Fig. 4.** Interpreted structure in the Norumbega restraining bend. $A =$ away, $T =$ toward. Redrawn from Swanson (1999) with his permission.

**Fig. 5.** Sketch defining symbols used in Eqs. (1)–(4). A mineral assemblage was initially at depth $h_b$ but is now exposed at the surface. At the time this mineral assemblage crystallized, the surface was $\Delta H$ meters higher than the present surface. Rock uplift has been $h_b - \Delta H$. Vertical dimension not to scale.
several variables affecting erosion rate that is included in most erosion laws, any law purporting to describe erosion rates over such time spans is only a rough approximation to reality. Despite this, a mathematical representation of this reality is useful in many studies.

Four forms of erosion law have been posited. Together with their solutions for $H(t)$, they are:

**Linear**:
$$E = v_H H \quad H(t) = H_0 e^{vt}$$

**Tangent**:
$$E = \frac{v_H H}{1-(H/H_c)} \quad t = \frac{6}{v} \left[ \ln \frac{H(t) + \frac{H_0^2}{2H_c^2} - (H(t)^2)}{H_c} \right]$$

**Power**:
$$E = v_H H^u \quad H(t) = \left[ H_0^{1-u} + \frac{1}{u} (1-u) \frac{v}{q_c} t \right]^{(-1/u)} \quad u \neq 1$$

**Exponential**:
$$E = v_H e^{qt} \quad H(t) = \frac{1}{q} \left( e^{-qH_0} - \frac{1}{6} v^{-1} q t \right)$$

Here, $v$ is a coefficient of erodibility, presumably dependent principally on lithological characteristics such as mineralogy or fracture density (e.g., Hooke and Rohrer, 1977); $t$ is the time in the past; $H_c$ is an empirical limiting value; and $u$ and $q$ are empirical constants. The subscripts $v$, $t$, $p$, and $e$ on $v$ refer to the respective laws. The linear and exponential laws are based on erosion rate and sediment yield data spanning only decades (e.g., Ahnert, 1970; Summerfield and Hutton, 1994), whereas the power and tangent laws have been tested using long-term data based on thermochronometry (Montgomery and Brandon, 2002). Note that the tangent law cannot be solved explicitly for $H(t)$. (Sources of these laws are discussed in Appendix C.)

In the original versions of Eq. (7), $H$ is defined locally (for example, as the mean difference in elevation between ridge crests and valley bottoms, with symbol $R$). As noted, however, the nature of our data constrains us to define it ‘globally’ as the mean elevation above (present) sea level. We justify this approach for our purposes because (i) sloughing or landsliding on slopes at and above the angle of repose is the primary means of delivering sediment to streams (Belmont et al., 2011); (ii) the frequency of such sloughing events depends on the rate at which sediment is removed from the base of the slope (Hooke, 2000; Montgomery and Brandon, 2002, p. 487); (iii) stream power ($= QS$ where Q is discharge and $S$ is channel gradient) is the best available measure of this removal rate; and (iv) in mature topography (without plateaus) a constant horizontal distance from a baselevel, elevation above sea level is better than local relief as a measure of stream power.

Sea level was not constant during this period, so $H$ was changing. The Matinicus Basin was bobbing up and down like a cork (Gibling et al., 2009). Periods of emergence are also indicated by erosion surfaces dating from the Mesozoic and from the Cenozoic that underlie the Gulf of Maine (Oldale and Uchupi, 1970; Poag and Schlee, 1984). However, to the extent that eustatic changes in sea level were small compared with $H$, they would not have affected erosion rates significantly. We found that incorporating sinusoidal oscillations in sea level in some of our models resulted in nominal changes in calculated values of $v_p$ but did not alter our overall conclusions.

### 3.5. Recapitulation I

We have defined $H(t)$ as the mean elevation of the land surface above modern sea level at some time, $t$, in the past and $\Delta H$ as the change in mean elevation resulting from removal of $h_0$ meters of crustal material. Our key assumptions so far have been:

- Airy isostasy is assumed. Where the crust was subjected to vertical forces from mantle convection or other sources, these forces would also need to be taken into consideration in a more refined analysis.
- We neglect $E_o$ as it is normally small.
- We also assume, initially, that $c = 0$.

### 4. Analysis

We know $H_p$ at 64 sites and have a good estimate of $h_0$ for these sites at a particular time, $t$. Thus we can calculate $H$ at that time (from Eqs. (1) and (2)). We also have the CBT time series of temperature.

To study patterns in these data, we first use our geobarometric data set to estimate the Paleozoic geothermal gradient at the Casco Bay site. This gives us a time series of $h_0$ at this site. We then consider the form of the erosion laws and also the influence of tectonic events on evolution of the landscape.

#### 4.1. Temperature gradient at the Casco Bay site

Sixteen of our geobarometric data sites on the northwest side of the Norumbega fault zone are within ~100 km of the Casco Bay site. Two lie ~39 km southwest of it (Fig. 1) and have an age of 293 Ma. The remaining 14 are between ~38 km northeast and ~83 km northwest of the site. Eleven of these are from samples dated to ~324 Ma; the remaining three are dated to 293, 386, and 399 Ma. The oldest CBT age from the northwest side of the Norumbega fault zone at the Casco Bay site is 284 Ma. Thus, we used the power erosion law to estimate pressures at 284 Ma for the 14 older sites. (As discussed later, the power law has advantages over the others, and we thus chose it for this step.) We then fit a variety of mathematical surfaces to all 16 pressures (Table 1) and used these surfaces to predict the pressure on the northwest side of the fault zone at 284 Ma. The pressures from the various models ranged from 3.7 to 5.3 kb. We discarded three models with RMS errors >0.8 kb and used the average of the remaining seven: 3.85 ± 0.14 kb. (The ± 0.14 kb uncertainty is based on the standard deviation of these seven estimates, not on the RMS error in prediction of the various surfaces.) Assuming a crustal density of 2750 kg/m$^3$, this pressure corresponds to a depth of 14.3 ± 0.5 km. This is an estimate of the depth, at 284 Ma, of rocks presently exposed at the surface. From the CBT temperature data we know the temperature at 284 Ma: 480 °C (hornblende). Allowing for a mean annual surface temperature of 15 °C, the resulting temperature gradient is 32.6 ± 1.2 K/km.

For comparison, the Late Paleozoic (~275--150 Ma) geotherm at Mt. Washington, 95 km NW of Casco Bay, was ~33 ± 7 K/km. This estimate is based on apatite fission track (AFT) (Roden-Tice et al., 2012) and 40Ar/39Ar (Eusden and Lux, 1994) analyses of samples from the summit and base of the mountain. The 40Ar/39Ar analyses yield the time at which the temperature passed through 320 °C, and the AFT analyses

<table>
<thead>
<tr>
<th>Interpolation method</th>
<th>Pressure, kb</th>
</tr>
</thead>
<tbody>
<tr>
<td>Global polynomial, order 1</td>
<td>3.71 ± 0.79</td>
</tr>
<tr>
<td>Local polynomial, order 1</td>
<td>3.71 ± 0.77</td>
</tr>
<tr>
<td>Krige, no detrending</td>
<td>3.74 ± 0.79</td>
</tr>
<tr>
<td>Radial basis function</td>
<td>3.83 ± 0.78</td>
</tr>
<tr>
<td>Inverse distance weighted, 0.8 smooth, power 2</td>
<td>3.88 ± 0.77</td>
</tr>
<tr>
<td>Inverse distance weighted, 0.8 smooth, power 3</td>
<td>3.98 ± 0.75</td>
</tr>
<tr>
<td>Krige, 1st-order detrending</td>
<td>4.07 ± 0.77</td>
</tr>
<tr>
<td>Local polynomial, order 2</td>
<td>4.60 ± 1.03</td>
</tr>
<tr>
<td>Global polynomial, order 2</td>
<td>4.67 ± 0.94</td>
</tr>
<tr>
<td>Krige, 2nd-order detrending</td>
<td>5.28 ± 1.00</td>
</tr>
</tbody>
</table>

* Uncertainties are RMS error in prediction of pressures at the 16 sites where pressure is known from geobarometric data.
yield the time when the temperature dropped below \( -90 \pm 10 \, ^\circ C \). By interpolating between the \(^{40}\text{Ar}/^{39}\text{Ar}\) AFT ages from the summit, one can estimate the temperature at the summit at the time the base was at \( 320 \, ^\circ C \) (274 Ma) and from this calculate a gradient (29 \( \pm \) 8 K/km).

A similar calculation using the data from the base yields a gradient at 152 Ma (36 \( \pm \) 13 K/km). The mean is 33 \( \pm \) 7 K/km. The uncertainty is calculated assuming uncertainties of \( \pm 10 \, ^\circ C \) in the muscovite closure and AFT temperatures, using published uncertainties in the ages and using standard rules of error propagation. Roden-Tice et al. (2012) estimated a geotherm of 36 K/km between 275 and 150 Ma, and from their AFT track-length modeling alone, 43 \( ^\circ C \)/km.

To compare these estimates with the present geothermal gradient in the Casco Bay area, we plotted the locations of 53 geothermal fluxes reported by Decker (1987). We converted these to gradients, using thermal conductivities of 2.4 Wm\(^{-1}\)K\(^{-1}\) in areas of granitic terrane and 2.9 Wm\(^{-1}\)K\(^{-1}\) in areas of amphibolite (Clauser and Huenges, 1995), and contoured 14 points within 80 km of the Casco Bay site. This yielded a gradient of 21 K/km at the Casco Bay site.

We thus assumed that the gradient was 32.6 K/km at 284 Ma, that it had been decreasing linearly with time since 380 Ma, and that it would continue to decrease linearly to 21 K/km. We suspect this decrease is largely owing to thickening of the crust (by 10–12 km) by vertical strain associated with the formation of the flower structure at the restraining bend in the Norumbega fault zone. Therefore we assumed that the gradient reached 21 K/km by 150 Ma. We used this gradient history to convert the CBT temperature data to depths, and then used Eqs. (1) with (2) to obtain \( H \). In Fig. 6A we plot \( H \) as a function of age for samples on either side of the Norumbega fault zone. For our purposes, any errors in the CBT data, with the possible exception of temperature histories based on AFT track length analyses, are likely negligible. Smooth curves can be drawn through the points up to the mid-Mesozoic. This suggests that conditions were relatively uniform during this time period.

4.2. Tests of erosion laws

We next explore the four erosion laws using this time series of \( H \). In Fig. 6A, the Mesozoic and Cenozoic AFT data (open symbols) disrupt what otherwise would be smooth curves terminating at the present heights of \( \sim 15 \, m \) (Fig. 6B). This disruption is a product of the Alleghanian orogeny and is discussed later. For the moment we ignore it noting that if the Alleghanian had not occurred, the topography would, by now, have eroded down to close to the present level. Thus to study the characteristics of the four erosion laws we plot, in Fig. 6B, only the other six points for the southeast side of the fault zone. We then fit least-squares curves through these six points for each of the erosion laws. With the exception of the linear law, the curves must be fit by trial and error techniques; we wrote FORTRAN code to do this.

The linear law is the most tractable, mathematically. However, it seriously overestimates the erosion rate when \( H \) is high (Fig. 6B). Montgomery and Brandon (2002, p. 487) found that it had the opposite
tendency and inferred that this was because mass movement processes, like landsliding, are more active in areas of high relief. Because such new erosional processes come into play as \( H \) increases, erosion rates are expected to increase more rapidly than linearly with increasing \( H \) (e.g., Whipple and Tucker, 1999).

Of the four erosion laws, the linear one consistently provides the worst fit.

The exponential law slightly underestimates \( H_p \). More importantly, \( dh/dt \) does not approach zero asymptotically as \( H \to 0 \), as expected, so predicted present-day erosion rates are too high.

The tangent law provides a reasonably good fit, although its prediction of \( H_p \) is high. Because it cannot be solved for \( H(t) \), it is inconvenient to use.

The power law commonly gives the best fit in this case and in many other permutations of the data that we have studied. Henceforth we focus on it.

4.3. Evaluating the constants in the power law

A puzzling characteristic of the power-law curve in Fig. 6B is that the exponent, \( u \) (Eq. (7c)) is \(-0.7\). As noted, nonlinear erosion laws, like the power law, are attractive because they allow a nonlinear increase in erosion rate as \( H \) increases. A value of \( u < 1 \) implies a nonlinear decrease instead. In Fig. 6A (solid curves), the value of \( u \) for the power-law fit to the Devonian-Jurassic data from the southeast is also \(< 1.0 \) (0.7) although that for the NW is not (1.1).

We inferred above that crustal thickening during formation of the flower structure was responsible for the decrease in geothermal gradient since 284 Ma. Homogeneous vertical stretching (thickening) would have increased the distance between rocks at different temperatures, reducing the temperature gradient and hence the cooling rate. In addition, the thickening implies \( c > 0 \) in Eq. (5) during this time span, leading to a spuriously low value of \( u \). If Airy isostasy prevailed, the thickening would have slowed the reduction in \( H \). Suffice it to say that we think the combined effect resulted in anomalously low slopes of curves of \( H \) versus \( t \) (Fig. 6A).

To evaluate \( u \), we turn instead to our full geobarometric data set. For each of our locations we calculated the temporal mean erosion rate, \( E \), from \( E = h_b/T \), where \( T \) is the time elapsed between crystallization of the mineral assemblage and the present. Then, using Eqs. (7c) and the definition of a temporal mean, we obtain,

\[
E = \frac{1}{T} \int_0^T E(t) \, dt = \frac{1}{T} \int_0^T v_p h^3 dt = \frac{1}{T} \int_0^T \left[ H_0^{1-u} + \frac{1}{b} (1-u) v_p t \right] \, dt
\]

Evaluating the integral and rearranging,

\[
v_p = \frac{6}{(1-u)T} \left[ \left( H_0 + \frac{TE}{b} \right)^{1-u} - H_0^{1-u} \right]
\]

But by definition,

\[
H_p + \frac{1}{b} TE = H_t
\]

where \( H_t \) is the initial height of the Paleozoic landscape, based on the equilibrium pressures. (Owing to isostatic uplift, the change in \( H \) is \(1/6\) of the total erosion, \( TE \).) So,

\[
v_p = \frac{6}{(1-u)T} \left( H_0^{1-u} - H_p^{1-u} \right)
\]

If \( v_p \) did not vary from site to site, we could solve for \( u \) (by trial and error) using values of \( H_0, H_p, \) and \( T \) from two different sites (subscripts 1 and 2), thus

\[
\frac{1}{T_1} \left( H_0^{1-u} - H_{p1}^{1-u} \right) = \frac{1}{T_2} \left( H_0^{1-u} - H_{p2}^{1-u} \right)
\]

However, \( v_p \) is an erodibility coefficient, dependent on lithology, so it will vary among sites. Furthermore, two clearly irrelevant solutions satisfy Eq. (10): \( u = 1 \) and \( u \to \infty \). Finally, one can readily verify that if \( (H_{t1} - H_{p1}) > (H_{t2} - H_{p2}) \), then \( (H_{p2}^{1-u} - H_{t2}^{1-u}) \) will remain greater than \( (H_{p1}^{1-u} - H_{t1}^{1-u}) \) as \( u \) increases from 1. Thus, for any two sites, a meaningful solution to Eq. (10) is only possible if \( T_1 < T_2 \) when \( (H_{t1} - H_{p1}) > (H_{t2} - H_{p2}) \). As a consequence, of the possible 2016 pairs of sites from our data base of 64 sites, only 140 yield meaningful values of \( u \) (Fig. 7).

The median value is 1.186 ± 0.14, which we round to 1.2. Once \( u \) is established, we use Eq. (9) to obtain \( v_p \) for each of our 64 sites. The median value is 0.0106 m\(^1\) yr\(^{-1}\), but this is meaningful only as a reference value as tectonic processes, in particular the Alleghanian collision, are not taken into consideration. Rounding, we use

\[
E(t) = 0.011H(t)^{1.2}
\]

where \( E \) is in meters per million years, \( H \) is in meters, and \( t \) is millions of years before present.

An \( H-t \) path with \( u = 1.2 \) is shown as a dashed line in Fig. 6A. By setting \( c = \xi_cH \) in Eq. (5), where \( \xi_c \) is the vertical strain rate resulting from the horizontal compression that produced the flower structure, we calculated the time series of \( \xi_c \) needed to produce the two observed \( H-t \) paths (Fig. 6C). These strain rates are reasonable: In the western U.S., \( \xi_c \) is typically between \(-1 \times 10^{-15} \) and \(+1 \times 10^{-15} \) s\(^{-1}\) (Platt et al., 2008, Fig. 5c).

5. Discussion

5.1. Setting the stage

In the Late Devonian (~380 Ma), the land surface upon which we now live was buried beneath up to ~20 km of rock. Making use of our assumption of Airy isostasy, the second of Eq. (7c) with \( u = 1.2 \), and the value of \( v_p \) for the individual sites obtained from Eq. (9), we normalized calculated topographic heights from sites with crystallization times between 365 and 420 Ma to 380 Ma and contoured these to obtain a glimpse of the topography at that time (Fig. 8A). A mountain range with a mean height of ~2000 (±400 m) appears to have then extended into western Maine. The uncertainty is based on an estimated uncertainty of ±0.65 kb in pressure. The highest peaks were probably in excess of 3000 m—perhaps comparable to today’s Rocky Mountains. Two geobarometric pressures (Appendix A: Mt. Waldo and Belfast) that appear to be robust suggest that high elevations extended eastward to the vicinity of the present Penobscot Bay. Northern and eastern Maine were lower, perhaps comparable to the present White Mountains of New Hampshire.

Our reconstructed 380 Ma topography suggests that sediment eroded from the north side of this Acadian mountain range would have been transported northeastward to eastern Maine and thence eastward to New Brunswick. This is consistent with the presence of thick sections of Carboniferous terrestrial clastic sediment around the edges of the Maritimes Basin in eastern New Brunswick (Currie, 1987) and northeastern Nova Scotia and of up to 12 km of Mid-Devonian to Early Permian sediment in the center of the basin, beneath the present Gulf of St. Lawrence (Murphy and Keppie, 1987; Gibling et al., 2009). At least during the Pennsylvanian and early Permian, paleocurrent data support such a southwesterly source (Gibling et al., 1992). Some of these beds may have extended into central Maine (Ryan and Zentilli, 1993, Fig. 7) and perhaps farther south (Gibling et al., 1992, p. 345);
but lacking preserved evidence for these, we have assumed that they were of negligible thickness.

Following the Alleghanian collision (~320 Ma), Maine was in the middle of the Pangea supercontinent. Our topographic reconstruction for this time (Fig. 8B) suggests that the Acadian mountains had now lost an average of ~3600 m of rock, reducing their height by ~600 m if fully compensated by isostatic uplift. However, to the south a new, somewhat higher range had resulted from the Alleghanian orogeny. Quite logically, Holdaway et al. (1988, p. 42) suggested that the rocks at the surface of this range were likely volcanic.

As rifting began in the Mesozoic (~220 Ma), block-faulted basins appeared to the southeast of these mountains that, by this time, had likely been reduced to less than half of their original height (Fig. 8C). These grabens now underlie the Bay of Fundy and the Gulf of Maine (Hutchinson et al., 1988) and contain up to 10 km of subaerial clastic sediment (Wade et al., 1996), probably largely derived from the southeastern slope of the Alleghanian mountains.

We estimate that Maine has lost an average of ~10 km of rock since 380 Ma (a volume of ~800,000 km³), reducing its mean elevation by ~1.6 km. Of this, ~7 km (or 560,000 km³) had vanished by 200 Ma.

5.2. Dip–slip motion on the Norumbega fault

The cooling (exhumation) curve (Fig. 6A) for the southeast side of the Norumbega fault zone prior to 100 Ma is shifted to the left with respect to that for the northwest side, implying that throughout this time span the topography on the northwest side was higher. At 284 Ma we estimate that the elevation difference was ~1100 m. The slopes of the lines through the points at any given time also differ, with the higher slope (= higher erosion rate) consistently on the northwest side. This would be expected if elevations were higher.

As most of the strike–slip motion along the fault zone is believed to have occurred prior to 360 Ma (Ludman et al., 1999), rocks on opposite sides may well have been in roughly the same relative horizontal position as they are now. The elevation difference thus suggests the presence of a range–bounding fault comparable to those in the Basin and Range Province of the western U.S. today. (Our geobarometric data in southeastern Maine (Fig. 8) are not spatially dense enough to define this scarp.) If it existed, however, this scarp was apparently restricted in extent as closure temperature data show no thermal discontinuities across the fault zone to the northeast and southwest of the restraining bend in at least the last 150 m.y. (West et al., 2008).

West et al. (1993, p. 1486) estimated, conservatively, that at 250 Ma, rocks presently exposed at the surface on the northwest side of the fault zone were ~120 K hotter than those on the southeast side. Assuming a geotherm of 30 K/km, they thus argue for ~4 km of west-side up displacement since 250 Ma. In our interpretation (Fig. 6A), at 284 Ma the temperature difference was already ~200 K. With our temperature gradient of 32.6 K/km at this time, this implies ~6 km of displacement since then. The CBT AFT data (Fig. 6A) suggest that rocks on either side of the fault zone have been at the same temperature since ~80 Ma, so vertical movement must have ended by then. West et al. (1993, p. 1485) pointed out that such estimates assume that thermal surfaces were essentially horizontal.

The difference in H across the fault at 284 Ma must have required several tens of millions of years to develop. The transition from predominantly dextral motion and thrusting, resulting in the flower structure, to extension with an increased dip–slip component is marked by a change from east-plunging lineations to southwest-plunging lineations on southeast-dipping shear planes (Swanson, 1999, pp. 92–94). This transition likely occurred in the Late Devonian (M. Swanson, University of Southern Maine, written communication 2012), so this may be when the scarp began to form. This extends West and Roden-Tice’s (2003) estimate of the time span of west-side up displacement back some tens of millions of years and increases the total offset several kilometers.

5.3. Decreased cooling rate during the early Mesozoic

The rate of cooling of rocks on both sides of the Norumbega fault zone in the Casco Bay area decreased going into the Mesozoic (Fig. 6A). The nearby Sebago batholith (roughly outlined by the 330 Ma contour in Fig. 1) appears to have had a quantitatively similar cooling history. This is based on the currently accepted intrusion age (~293 Ma; Tomascak et al., 1996) and temperature (~650 °C; Aleinikoff et al., 1985) of the Sebago and on two monazite ages and other cooling data from the nearby White Mountain batholith (Aleinikoff et al., 1985, Fig. 4). We consider three possible causes of this decrease.

(i) An intrusion at about mid-depth in the crust could have decreased the cooling rate. The large Sebago batholith lies ~5 km west of Casco Bay, but it is almost 100 m.y. too old. Modeling of the thermal response to this thin intrusion suggests that the effects would have dissipated in ~10 m.y. (De Yoreo et al., 1989). There are, however, several smaller plutons and hundreds of Late Cretaceous maﬁc dykes, ranging up to ~75 m in width, in this part of Maine (Poland and Faul, 1977; McHone and Butler, 1984; Hussey et al., 2008; West et al., 2008, Fig. 3). Again, though, our modeling of the thermal perturbation caused by an 1100 °C basalt sill, 100 m thick, intruded ~1 km beneath
the samples suggests that the sill would have cooled quickly and, after 40,000 years, would have been only ~15 °C warmer than the country rock at the time of intrusion. Thermal perturbations exceeding 25 °C were confined to within a kilometer of the sill. Thus, it seems unlikely that intrusions of the scale observed at the present surface could be responsible for the observed temperature history.

(ii) The cooling rate could also have been decreased by an increase in the geothermal gradient conducting heat upward from below to the then level of the samples, accompanying stretching and thinning of the crust during Mesozoic rifting. The thinning was probably <7 km (Unger et al., 1987), however, as the actual rifting took place farther southeast. Seven kilometers of thinning would have increased the geothermal gradient ~4 K/km, which again is not enough, alone, to have significantly decreased the cooling rate.

(iii) In terms of topography, the decrease in cooling rate implies a decrease in erosion rate while the topography was still ~600 m above base level. Such a decrease in erosion rate is a logical consequence of the Alleghanian collision, which placed presently coastal Maine in the middle of Pangea, hundreds of kilometers from the ocean. We think this is why the cooling rate decreased.

5.4. Delayed cooling during the Cretaceous—scarp retreat?

The Cretaceous AFT data suggest rather abrupt increases in erosion rate, first on the southeast at ~140 Ma and then on the northwest
–40 m.y. later (dashed curves in Fig. 6A). In terms of temperatures, the curves suggest that samples on the northwest remained above –90 °C for 40 m.y. longer than those on the southeast. Intrusions at the right time and place could be responsible for this. However, we favor an explanation involving rifting.

Rifting would have been preceded by stretching and thinning of the crust. One likely response to thinning would be lowering of the land surface accompanied by a rise in the Moho. Upwelling of hot mantle material, however, likely inhibited lowering, while the stretched-but-still-connected crust prevented full response to the upward forces arising from upwelling. Upon failure, rift flanks tend to bounce upward rather suddenly (e.g., Chorowicz, 2005; Sachau and Koehn, 2010). This is variously attributed to upwelling, to thinning, or to a combination of the two (Turcotte and Emerman, 1983).

Once the crust began to fail, graben faulting accompanied by flank uplift would have produced escarpments. Erosion rates on the uplifted plateau-like surfaces would have remained low, thus extending the period of slow cooling. High erosion rates on the scarp faces, however, would have initiated scarp retreat to the northwest. We speculate that a scarp bordering a graben on the inner edge of the Gulf of Maine first reached the site on the southeast side of the fault zone, and then, ~40 Ma later, the site on the northwest side. As the sites are ~17 km apart, this implies a retreat rate of ~0.4 mm a\(^{-1}\). This is a reasonable rate. Cosmogenic nuclide studies suggest that the rate of retreat of the Drakensberg escarpment in South Africa is about 0.1–0.2 mm a\(^{-1}\) (Fleming et al., 1999; Brown et al., 2002), and the present location of the Great Escarpment of southeastern Australia (Ollier, 1982) relative to its presumed initial location and age suggests a rather high retreat rate of 2 mm a\(^{-1}\) (Seidl et al., 1996).

Delayed cooling of the Biddeford pluton (West et al., 2008, p. 293, Fig. 3), ~50 km southwest of the Casco Bay site, may also be a consequence of this particular scarp retreat. A sample on the southeast side of the pluton cooled through 100 °C at 147 Ma, while one 10.6 km to the northwest did not do so until 100 Ma, suggesting a retreat rate of 0.23 mm a\(^{-1}\).

Finally, the foothills of the western mountains in Maine run roughly parallel to the coast and ~60 km inland from the Norumbega fault zone in the Casco Bay region. If these foothills are a remnant of the escarpment, the retreat rate has been ~0.6 mm a\(^{-1}\).

The continental margin, the boundary between continental and oceanic crust, is ~400 km southeast of Casco Bay. If the abrupt increases in erosion rate suggested by the AFT temperature histories are, indeed, a result of scarp retreat, it seems likely that this particular escarpment was initiated along the inner edge of the Gulf of Maine rather than at the more distant continental margin. Otherwise, the initial erosion rate would have had to have been unrealistically high. However, escarpments likely bordered the grabens now submerged beneath the Gulf of Maine (Ballard and Uchupi, 1972; Uchupi and Bolmer, 2008), as they now bound the East African Rift valley—the latter commonly exceeding 1000 m in height (Chorowicz, 2005).

5.5. Cretaceous and Cenozoic cooling

Following this Mesozoic rejuvenation, \(H\) might be expected to again decline along a parabolic pathway similar to those in the Paleozoic. Such a path is shown by the dotted line in Fig. 6A. However, the AFT track length simulations suggest, instead, a flattening at 80 Ma, and then a steepening again at ~35 Ma. The steepening coincides rather well with a late Eocene unconformity in the stratigraphy of Georges Bank, near the edge of the continental shelf, and elsewhere along the eastern seaboard (Poag and Schlee, 1984; Uchupi and Bolmer, 2008, pp. 40–41), and with a probable sharp drop in sea level of possibly as much as 400 m at ~30 Ma (Vail et al., 1977). The high stand, which lasted from the Cretaceous to the mid-Oligocene (30 Ma), would have inhibited erosion and thus delayed cooling, while the lowered sea level would have initiated renewed erosion.

5.6. Rock erodibility and erosion rates

During Unroofing, erosional agents would have encountered a variety of lithologies—initially sedimentary but later metamorphic and plutonic—with varying resistance to erosion. At the sites for which we have geobarometric data, total erosion varied from 1 to 23 km and averaged 12 km. As many plutons are believed to be tabular bodies, only 1 to 2 km thick (De Yoreo et al., 1989, p. 180; Hogan and Sinha, 1989, p. 26), more than one pluton may have been encountered in some places. Clearly, we cannot know the sequence of lithologies that were eroded away at any given place, nor should we expect a close correlation between erosion coefficients and the lithologies presently exposed.

Apparently, however, \(\nu_p\) increases systematically from north to south (Fig. 9). There are three likely explanations for this pattern. First, given the topography that we visualize (Fig. 8), it is likely that throughout much of the past 400 m.y. stream power (QS) was lower in the interior of Maine than it was along the present coast. Secondly, the Alleghanian collision probably resulted in some vertical extension in the crust some distance northwest of the (likely volcanic) Alleghanian mountain range itself. Any such strain that occurred above the then level of the sample that yielded the geobarometric data would have increased the amount of erosion necessary to expose the sample. Because this was not taken into account in our calculations, our estimate of \(\nu_p\) would be too low. Thirdly, the steep gradient in \(\nu_p\) nearer

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**Fig. 9.** Spatial distribution of \(\nu_p\). Interpolation was by kriging in Arcmap followed by smoothing in Adobe Illustrator. Contour interval: 0.0025 m a\(^{-2}\) a\(^{-1}\).
Post Acadian erosion has removed ~8 × 10^5 km³ of rock from the state. With the use of our full geobarometric data set, we obtained the value of features. Assuming Airy isostasy, we then calculated at the times at which they passed through certain Ar closure temperature 10 °C at present. This enabled us to estimate depths to the CBT samples mean annual surface temperature decreased from 15 °C at 284 Ma to 150 Ma and remaining constant thereafter, and that, similarly, the (21 K/km). We assumed that the gradient decreased linearly with the coast may reflect accelerated erosion on the postulated retreating escarpment following break up of Pangaea. In order to compare our erosion rates with those of others, we calculated H(t) from the second of Eq. (7c) with H_p = 200 m, the median present elevation of our geobarometric sites. We then calculated E(t) from Eq. (11) for this hypothetical site. With the caveats that the resulting curves are fairly sensitive to the value we chose for an present elevation of our geobarometric sites. We then calculated E(t) for this hypothetical site. For these values of H_p and at that any given time H varied across New England, our erosion rates are consistent with those obtained by studies in nearby areas (Fig. 10).

5.7. Recapitulation II

We obtained two estimates of the geothermal gradient in the vicinity of the CBTE site, one at 284 Ma (32.6 K/km) and one at present (21 K/km). We assumed that the gradient decreased linearly with time, passing through 32.6 °C/km at 284 Ma to become 21 °C/km at 150 Ma and remaining constant thereafter, and that, similarly, the mean annual surface temperature decreased from 15 °C at 284 Ma to 10 °C at present. This enabled us to estimate depths to the CBT samples at the times at which they passed through certain Ar closure temperatures. Assuming Airy isostasy, we then calculated H at these times. A plot of H versus elapsed time since 350 Ma revealed several idiosyncrasies for which we have offered explanations.

6. Conclusions

• Given the uncertainties involved, any one of the four erosion laws would likely provide as good an approximation to reality as is possible with presently available data. However, the power law, E(t) = v(t)H(t)^β combines mathematical simplicity with a nonlinear variation with altitude and an asymptotic approach to zero as H → 0. For these reasons we prefer it.
• The value of v derived from the Casco Bay H–t curves is anomalously low. This is likely due to crustal thickening during the Early Paleozoic.
• With the use of our full geobarometric data set, we obtained u = 1.2.
• The Acadian orogeny produced mountains, perhaps comparable to the Rockies of today, in western Maine at ~380 Ma. By 320 Ma, these low. This is likely due to crustal thickening during the Early Paleozoic.
• Post Acadian erosion has removed ~8 × 10^5 km³ of rock from the state. During the initial phases of this unroofing, the eroded sediment was deposited in the Maritimes Basin, now mostly submerged beneath the Gulf of St. Lawrence. Later, it also accumulated in grabens now beneath the Bay of Fundy and Gulf of Maine.
• The persistently higher temperature of rocks on the northwest side of the Norumbega fault zone from Permian through Early Cretaceous suggests dip–slip displacement on the fault during that time, with well over 6 km of northwest-side up offset. This extends West and Roden-Tice's (2003) estimate upward in magnitude and backward in time.
• In the Mesozoic, the H–t paths for opposite sides of the Norumbega fault zone unexpectedly asymptotically approach ~600 m. This likely reflects suturing of Laurentia and Gondwana to form Pangaea, putting Maine in the middle of a vast supercontinent.
• The increase in slope of the H–t curves in the Cretaceous suggests a sudden increase in erosion rate. We attribute this to rifting during early stages of opening of the Atlantic. We attribute the time delay of ~40 m.y. across the Norumbega fault zone to retreat of a rift-flank escarpment. A similar abrupt increase in slope at 40 Ma may reflect a substantial fall in sea level at that time.
• The highest values of v_p appear to have been along the southwestern coast, perhaps reflecting generally steeper stream gradients throughout the past 400 m.y.
• Thermochronologic and geobarometric data are a trove of information on paleotopography, but interpretation of such data in terms of likely geomorphic and tectonic evolution is commonly relegated to a secondary status. Consideration of the implications of these data in terms of topography may provide insights that would otherwise escape detection. Topographic evolution exerts a first-order influence on cooling paths and can serve as a starting point for analysis.

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Appendices. Supplementary data

Supplementary data to this article can be found online at http://dx.doi.org/10.1016/j.geomorph.2013.12.015.

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