I wish to thank the AE for his open-mindedness in allowing me to submit a major revision to the discussion paper. I apologize for the time it has taken me to make the modest changes he has asked for. I promise that any future revisions will be completed more promptly. I have a strong incentive to get this paper published, in part because I do not wish the flawed discussion paper to remain the paper of record of my work on this topic.

Q: "One of the main issues concerns the computation of what Pelletier calls residuals and is referred to as "computing multiple intercepts" from a single regression by several of the reviewers. Although I see what the author has been trying to do, I also appreciate the point made by the reviewers that the value of P0 obtained by regression depends of course on the assumed value for h0 used in equation (1). Pelletier should call his residuals P' or Pr, to avoid confusion and refrain from making the assertion that by analysing the residuals, he is providing further constraints on what controls P0, as defined by Heimsath."

A: Please note that the two values of h_0 I have adopted come from the Heimsath et al. analysis. I have not assumed any value of h_0. Rather, I have assumed that the value of h_0 is sufficiently uniform within the two areas of the study site identified by Heimsath et al. (2012) (i.e., those with slopes above and below 30 degrees) that we can be confident in the trends of $P_{0,resid}$ (now called Pr) that I have identified. I understand perfectly well the issue that the reviewers have raised. The reviewers are concerned that a significant amount of the variation that I am attributing to Pr values is instead due to spatial variations in h_0. However, in order for the relationship between Pr and slope to be significantly affected, h_0 would have to have a systematic dependence on slope. As I have already noted in my previous rebuttal, Heimsath et al. (2012) clearly demonstrated that this was not the case. These authors considered two end member slope regimes and found that the average h_0 values for these two regions differed by only 0.05 m. As I noted previously, this difference in h_0 values corresponds to a difference in Pr values that is more than 100 times smaller than the actual variation in Pr values for a soil 30 cm in thickness. More broadly, h_0 values have now been estimated for many sites in widely different climates and have found to vary at most by approximately a factor of 2, compared to variability in P_0 values of approximately 3 orders of magnitude. In the revised paper I have added the following paragraph on this point: "Estimating Pr values using the residuals of the regressions of Heimsath et al. (2012) assumes that h0 has sufficiently limited variation within the two subsets of the study site considered by Heimsath et al. (2012) (i.e., those with Sav values above and below 30°) that any such variation would not affect the conclusions of the paper. For example, in order for the relationship between Pr and Sav (i.e., Figs. 2A-2C) to be significantly affected by variations in h0, h0 would have to have a systematic dependence on Sav. For example, if systematically lower values of h0 occur at steeper slopes and this effect is not accounted for, the result could be a biasing of Pr values downward in such regions. Heimsath et al. (2012) clearly demonstrated that no such systematic dependence exists. These authors considered two end member slope regimes and found that the average h0 values for these two regions differed by only 0.05 m (0.32 m vs. 0.37 m). At a soil thickness of 0.3 m, this difference corresponds to Pr differences of approximately 10% (i.e., exp(-0.30/0.32) vs. exp(-0.30/0.37)). This difference is more than 100 times smaller than the variation in Pr values. The difference becomes even smaller for soils thinner than 0.3 m."

The stated basis for Heimsath and Whipple's disagreement with me was, in some cases, simply that they had looked for a correlation (e.g., with climate) and found none, so therefore none must exist. I respectively ask that they consider my work on its merits. I propose that I have made a useful advance in this paper. I have provided a process-based explanation (i.e., topographic-stress-induced opening of fractures) for the slope dependence on Pr in the SGM. My analysis is an alternative to the theory proposed by Heimsath et al. (2012) that erosion rates control potential soil production rates both in the SGM and globally (the latter based on correlations between erosion rates and potential soil production rates globally). I believe the

evidence demonstrates that slope, not erosion rate, controls potential soil production rates, and only in regions of significant compressive stress. I appreciate that erosion rate and slope are closely correlated in the SGM, but such a correlation does not hold everywhere. My alternative is a very different view from Heimsath et al. (2012) but is supported using the same data they presented. I believe that the availability of two clear alternative models will spur new investigations on this problem.

I am somewhat confused by the recommendation that I change the variable P_0 since I already changed it to $P_{0,resid}$ in my previous submission to make it clear that my analysis uses residuals to estimate the soil maximum or potential sol production rate. I take the AE's recommendation to mean that I cannot use P_0 in any form whatsoever (even modified to indicate the use of residuals). In response, I have changed the symbol from P_0 to Pr. I am concerned that this change will be confusing to readers since it is well established in the literature that P_0 refers to the soil production rate as soil thickness approaches zero, which is equivalent to the maximum soil production rate in the case of an exponential soil production function. I am not aware of any publication, including any by Heimsath et al., that defines P_0 as the y intercept of a fit to an exponential soil production function. Every time that P_0 (or its precursors $epsilon_0$ (Heimsath et al.) or W_0 (Furbish and Fagherazzi)) have been defined, it is as a property of nature (i.e., the soil production in the limit of zero soil thickness), not as a mathematical construct (the y-intercept of a regression). I believe that variables should always be defined as properties of nature rather than as the results of a particular method of data analysis. An analogy can be made here with the exponents m and n of the stream power law. These exponents are estimated in multiple ways, just as P_0 can be estimated in multiple ways. If we used a different symbol to represent the same property of nature each time a different method of estimation was used, the result would be chaos. In numerical models the values of m, n, or P_0 are often prescribed rather than estimated using data. In all models that I am aware of that include soil production, P_0 is used to represent the potential soil production rate regardless of how that variable is constrained (i.e., simply prescribed or fit to data). Such a use will, of course, no longer be allowed if this brand new definition of Heimsath is adopted as the only possible definition of P_0.

Q: "Pelletier argues now that the main factor controlling the residuals is tectonic stress. He shows that a simple relation obtained by Savage and Swolfs to predict stress can be used to predict the residuals. I note that in his revised version Pelletier makes references to Figure 2 whereas he means 3, I think. Pelletier also claims that there is an additional climatic control which he assesses by comparing the already corrected residuals with those further corrected by a climate correction factor. I am still somewhat confused on how that coefficient is determined."

A: I apologize for reversing the order of Figures 2 and 3 in my proposed revision. This has been corrected. I would be happy to clarify how the climate factor has been determined, but I would need more information from the AE on what sentences are confusing.

Q: "In the first version of the paper submitted to EsurfD, Pelletier argued for a control by fracture density which he estimated by introducing a damage index. He has revised not only the definition of the damage index, but also lessened his conclusion concerning the importance of damage/fracture density, arguing now for a control by topographic stresses alone. In this way, he has responded constructively to many of the reviewers criticisms."

A: Thank you. Please note that I have not redefined the damage index. The damage index in this revision is the same as the discussion paper. Please see AC6 for more on this issue.

Q: "Another major point of disagreement with the reviewer(s) is the measure of fit to data used by the authors (regression coefficient), whereas one of the reviewers argues for another, more appropriate measure, based on so-called Nash-Sutcliffe statistics. Pelletier argues that the measure proposed by the reviewer is inappropriate for assessing regressions. I am not an expert on this, but interestingly, the two methods yields the same fitness measure and Pelletier argues that it is appropriate."

A: There is not much more I can say on this point. As I argued in my earlier rebuttal, the N-S statistic does not apply to regression models (by definition).

Q: "All reviewers wonder why Pelletier has not performed a more classical multi variate analysis of the residuals, rather than privileging a step-by-step reduction of the residuals by incrementally incorporating potential processes through a simple mathematical description of the process. The main reason, I believe, of Pelletier's choice is that the relationships between variables (such as slope, climate, etc.) and residuals may not be linear at all and/or depend on thresholds which would be difficultly extracted from a simple cluster or multi variate analysis. I don't think, however, that this is clearly stated in the revised manuscript and it should (if my interpretation is correct) or an alternate explanation should be given."

A: I agree entirely with the AE that "the relationships between variables (such as slope, climate, etc.) and residuals may not be linear at all and/or depend on thresholds which would be difficultly extracted from a simple cluster or multi variate analysis." It was not a choice to forgo the multivariate linear fit to the data (with or without log transformation) proposed by reviewer 2. This approach would have been plainly inconsistent with the complex nonlinear relationships in the data, which my analysis and Heimsath et al.'s (2012) analysis clearly demonstrate are present. I have added the following paragraph to the discussion on this point: "This paper adopts a stepwise regression and cluster analysis approach that builds upon the regression analysis that Heimsath et al. (2012) used to characterize the dependence of soil production rate on soil thickness. Stepwise regression is a standard approach in statistics in which the residuals of a statistical regression are computed and additional controls tested for. I did not apply simultaneous multivariate linear regression (with or without log transformation) because such an approach would have been inconsistent with the complex nonlinear relationships in the data documented by Heimsath et al. (2012) and the analyses presented here."

I would like to make clear that I am not opposed to trying any additional analyses that the reviewers think are important. It is just not clear to me that an alternative approach has been suggested that is consistent with the complex nonlinear trends in the data nor is it clear to me why the approach I have taken has been judged to be inadequate.

Q: "There a few (unfortunate) comments by one of the reviewers that almost suggest that the author may have not properly reported the data (difference between a table and a diagram); the author has verified and confirmed that this is not the case. I checked a few points and saw no difference between table and plot. I will specifically ask the reviewers to refrain from making such comments in further reviews."

A: I had no problem with this comment. It seemed to me to be a minor, unintentional oversight by the reviewer (or by myself – I cannot say for sure that there was no discrepancy). I have much greater concern over the reviewer's accusation that my work is akin to data fabrication and his refusal to honor the GSA Ethical Guidelines for Publication in his role as a reviewer for a previous version of this paper. More generally, I am concerned about potential conflicts among some of the reviewers that may prevent a fair and impartial judgement of my paper. I respectfully ask the AE and Editor to consider these concerns.

Quantifying the controls on potential soil production rates: A case study of the San Gabriel Mountains, California

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Abstract. The potential soil production rate, i.e., the upper limit at which bedrock can be converted into transportable
material, limits how fast erosion can occur in mountain ranges in the absence of widespread landsliding in bedrock or intact
regolith. Traditionally, the potential soil production rate has been considered to be solely dependent on climate and rock
characteristics. Data from the San Gabriel Mountains of California, however, suggest that topographic steepness may also
influence potential soil production rates. In this paper I test the hypothesis that topographically induced stress opening of
pre-existing fractures in the bedrock or intact regolith beneath hillslopes of the San Gabriel Mountains increases potential
soil production rates in steep portions of the range. A mathematical model for this process predicts a relationship between
potential soil production rates and average slope consistent with published data. Once the effects of average slope are
accounted for, evidence that temperature limits soil production rates at the highest elevations of the range can also be
detected. These results confirm that climate and rock characteristics control potential soil production rates, but that the
porosity of bedrock or intact regolith can evolve with topographic steepness in a way that enhances the persistence of soil
cover in compressive-stress environments. I develop an empirical equation that relates potential soil production rates in the
San Gabriel Mountains to the average slope and a climatic index that accounts for temperature limitations on soil production
rates at high elevations. Assuming a balance between soil production and erosion rates at the hillslope scale, I illustrate the
interrelationships among potential soil production rates, soil thickness, erosion rates, and topographic steepness that result
from the feedbacks among geomorphic, geophysical, and pedogenic processes in the San Gabriel Mountains.

Keywords: soil production, cosmogenic radionuclides, topographically induced stress, San Gabriel Mountains

1 Introduction

The potential soil production rate (denoted herein by ~~P_0~~P_r) is the highest rate, achieved when soil cover is thin or
absent, that bedrock or intact regolith can be converted into transportable material at each point on Earth's surface. ~~P_0~~P_r
values are the rate-limiting step for erosion in areas where landsliding in bedrock or intact regolith is not widespread,

because soil must be produced before it can be eroded. Slope failure in bedrock or intact regolith is common in some fine-grained sedimentary rocks (e.g., Griffiths et al., 2004; Roering et al., 2005) but relatively uncommon in granitic rock types.

Despite its fundamental importance, the geomorphic community has no widely accepted conceptual or mathematical model for potential soil production rates. Pelletier and Rasmussen (2009) took an initial step towards developing such a model by relating ~~P_0~~P_r values in granitic landscapes to mean annual precipitation and temperature values. The goal of this model was to quantify how water availability and vegetation cover control the potential soil production rate across the extremes of Earth's climate. The Pelletier and Rasmussen (2009) model predicts ~~P_0~~P_r values consistent with those reported in the literature from semi-arid climates, where ~~P_0~~P_r values typically range from ~30-300 m/Myr. In humid climates, the Pelletier and Rasmussen (2009) model predicts ~~P_0~~P_r values greater than 1000 m/Myr (Fig. 2A of Pelletier and Rasmussen, 2009). This is broadly consistent with measured soil production rates of up to 2500 m/Myr in the Southern Alps of New Zealand where the mean annual precipitation (MAP) exceeds 10 m (Larsen et al., 2014). The Pelletier and Rasmussen (2009) model was a useful first step, but clearly not all granites are the same. In particular, variations in mineralogy (Hahm et al., 2014) and bedrock fracture density (Goodfellow et al., 2014) can result in large variations in soil production rates in granites within the same climate.

The San Gabriel Mountains (SGM) of California (Fig. 1) have been the focus of many studies of the relationships among tectonic uplift rates, climate, geology, topography, and erosion (e.g., Lifton and Chase, 1992; Spotila et al., 2002; DiBiase et al., 2010; 2012; DiBiase and Whipple, 2011; Heimsath et al., 2012; Dixon et al., 2012). These studies take advantage of a significant west-to-east gradient in exhumation rates in this range. Spotila et al. (2002) documented close associations among exhumation rates, mean annual precipitation (MAP) rates, and the locations and densities of active tectonic structures. Mean annual precipitation (MAP) rates vary by a factor of two across the elevation gradient and exhibit a strong correlation with exhumation rates (Spotila et al., 2002, their Fig. 10). Lithology, which varies substantially across the range (Fig. 1), also controls exhumation rates. Spotila et al. (2002) demonstrated that exhumation rates are lower, on average, in rocks relatively resistant to weathering (i.e., granite, gabbro, anorthosite, and intrusive rocks) compared to the less resistant schists and gneisses of the range (Spotila et al., 2002, their Fig. 9). This lithologic control on long-term erosion rates can control drainage evolution. For example, Spotila et al. (2002) concluded that the San Gabriel River has exploited

the weak Pelona Schist to form a rugged canyon between ridges capped by more resistant Cretaceous granodiorite (e.g., Mount Baden Powell). Spotila et al. (2002) concluded that landscape evolution in the SGM was controlled by a combination of tectonics, climate, and rock characteristics.

Heimsath et al. (2012) provided a millennial-time-scale perspective on the geomorphic evolution of the SGM. These authors demonstrated that soil production rates (P) and erosion rates (E) in rapidly eroding portions of the SGM greatly exceed P_r values in slowly eroding portions of the range. Heimsath et al. (2012) concluded that high erosion rates, triggered by high tectonic uplift rates and the resulting steep topography, cause potential soil production rates to increase above any limit set by climate and bedrock characteristics. Their results challenge the traditional view that P_r values are controlled solely by climate and rock characteristics.

Recent research, stimulated by shallow seismic refraction and drilling campaigns, has documented the importance of topographically induced stresses on the development of new fractures (and the opening of pre-existing fractures) in bedrock or intact regolith beneath hillslopes and valleys (e.g. Miller and Dunne, 1996; Martel, 2006; 2011; Slim et al., 2014; St. Clair et al., 2015). In this process, the bulk porosity of bedrock and intact regolith evolves with topographic ruggedness (i.e., topographic slope and/or curvature). In a compressive stress environment, topographically induced stresses can result in lower compressive stresses, or even tensile stresses, in rocks beneath hillslopes. As an elastic solid is compressed, surface rocks undergo outer-arc stretching where the surface is convex-outward (i.e., on hillslopes), reducing the horizontal compressive stress near the surface and eventually inducing tensile stress in areas of sufficient ruggedness. Such stresses can generate new fractures or open pre-existing fractures in the bedrock or intact regolith, allowing potential soil production rates to increase. In this paper I test whether potential soil production rates estimated using the data of Heimsath et al. (2012) are consistent with the topographically induced stress fracture opening hypothesis in the SGM. This hypothesis predicts a relationship between P_r values and average slope that is consistent with the data of Heimsath et al. (2012). Once the effects of average slope are accounted for, I test the hypotheses that climate, lithology, and local fault density also influence P_r values. I then use the resulting empirical model for P_r values to map the spatial variations in potential soil production rates, soil thickness, erosion rates, and topographic steepness across the range in order to illustrate the interrelationships among these variables.

2 Data analysis and mathematical modeling

2.1 Controls on potential soil production rates in the SGM

~~P_0 values~~ Estimates of the maximum or potential soil production rate (i.e., the soil production rate obtained when the buffering effects of soil, if present, are factored out of the measured soil production rate) for the SGM can be estimated using the residuals obtained from the regression of soil production rates to soil thicknesses reported by Heimsath et al. (2012) (their Fig. 3). The exponential form of the soil production function quantifies the decrease in soil production rates with increasing soil thickness:

$$P = \cancel{P_0 e^{-h/h_0}} P_r e^{-h/h_0}, \tag{1}$$

where h is soil thickness and h_0 is a length scale quantifying the relative decrease in soil production rates for each unit increase in soil thickness. Heimsath et al. (2012) obtained $h_0 = 0.32$ m for locations with an average slope, S_{av}, of less than or equal to 30° and $h_0 = 0.37$ m for locations with $S_{av} > 30°$. S_{av} is defined by Heimsath et al. (2012) as the average slope over hillslopes adjacent to each sample location. ~~P_0 values~~Values of the maximum soil production rate (Supplementary Table 1) can be estimated as the residuals obtained by dividing P values by the exponential term in equation (1):

$$\cancel{P_{0,resid}}P_r = \begin{array}{l} Pe^{h/0.32\,\text{m}} \ \ \text{if } S_{av} \leq 30° \\ Pe^{h/0.37\,\text{m}} \ \ \text{if } S_{av} > 30° \end{array}. \tag{2}$$

~~where $P_{0,resid}$ denotes P_0 values estimated using the residuals of the regression.~~ Note that equation (2) is equivalent to subtracting the logarithms of the exponential term from the logarithms of P values, since division is equivalent to subtraction under log transformation. Log transformation is appropriate in this case because P values are positive and positively skewed (i.e., there are many P values in the range of 50-200 m/Myr and a smaller number of values in the range of 200-600 m/Myr that would be heavily weighted in the analysis if the data were not log-transformed). ~~$P_{0,resid}$~~P_r values estimated from equation (2) increase, on average, with increasing S_{av} (Fig. 2A). ~~$P_{0,resid}$~~P_r values exhibit an abrupt increase at an S_{av} of approximately 30°.

Heimsath et al. (2012) did not include data points from locations without soil cover in their regressions because these data points appear (especially for areas with $S_{av} > 30°$) to fit below the trend of equation (1). This implies that a humped production function may be at work in some portions of the SGM. The mean value of P from areas with $S_{av} \leq 30°$ that lack soil cover is 183 m/Myr, i.e., slightly higher than, but within 2σ uncertainty of, the 170 ± 10 m/Myr value expected based on the exponential soil production function fit by Heimsath et al. (2012). As such, the evidence indicates that for areas with $S_{av} \leq 30°$, data from locations with and without soil cover are both consistent with an exponential soil production function. The mean value of P from areas with $S_{av} > 30°$ that lack soil cover is 207 m/Myr, i.e., significantly lower than the 370 ± 40 m/Myr expected based on the exponential soil production function. This suggests that a hump may exist in the soil production function for steep ($S_{av} > 30°$) slopes as they transition to a bare (no soil cover) condition. To account for this, I estimated $P_0$$P_r$ to be equal to 1.78P (i.e., the ratio of 370 to 207) at locations with $S_{av} > 30°$ that lack soil cover.

The SGM has horizontal compressive stresses of ~10 MPa in an approximately N-S direction at depths of less than a few hundred meters (e.g., Sbar et al., 1979; Zoback et al., 1980; Yang and Hauksson, 2013). The development of rugged topography can lead to topographically induced fracturing of bedrock and/or opening of pre-existing fractures in compressive-stress environments (e.g., Miller and Dunne, 1996; Martel, 2006; Slim et al., 2014; St. Clair et al., 2015). Given the pervasively fractured nature of bedrock in the SGM (e.g., Dibiase et al., 2015), I assume that changes in the stress state of bedrock or intact regolith beneath hillslopes leads to the opening of pre-existing fractures (i.e., an increase in the bulk porosity of bedrock or intact regolith) rather than the fracturing of intact rock. I adopt the analytic solutions of Savage and Swolfs (1986), who solved for the topographic modification of regional compressive stresses beneath ridges and valleys oriented perpendicular to the most compressive stress direction. Savage and Swolfs (1986) demonstrated that the horizontal stress (σ_{xx}) in bedrock or intact regolith becomes less compressive under ridges as the slope increases (Fig. 3). In landscapes with a maximum slope larger than 45˚ (equivalent to an average slope of approximately 27˚ or atan(0.5) in the mathematical framework of Savage and Swolfs, 1986), bedrock or intact regolith that would otherwise be in compression develops tensile stresses close to the surface beneath hillslopes (Fig. 3A). An average slope of 27˚ is close to the threshold value of 30˚ that Heimsath et al. (2012) identified as representing the transition from low to high P_0 values in the SGM. Therefore, the abrupt increase in $P_{0,resid}$$P_r$ values at approximately 30˚ is consistent with a transition from compression to tension in bedrock or

intact regolith beneath hillslopes of the SGM. In addition to this sign change in the horizontal stress state in the rocks beneath hillslopes of the SGM, the Savage and Swolfs (1986) model predicts a gradual decline in horizontal compressive stress as S_{av} increases between 0 and approximately 27° (Fig. 3B):

$$\frac{\sigma_{xx}}{N_1} = \frac{2 - 4S_{av}}{(2 + 4S_{av})(1 + 4S_{av})} \qquad (3)$$

5 where N_1 is the regional maximum compressive stress and S_{av} has units of m/m in equation (3). Equation (3) is simply equation (36) of Savage and Swolfs (1986) expressed in terms of the average slope from the drainage divide to the location of maximum slope rather than the shape parameter b/a used by Savage and Swolfs (1986). Note that the tangent of the slope angle (units of m/m) is averaged to obtain S_{av} in all cases in this paper. However, after this averaging, S_{av} is reported in degrees in some cases to facilitate comparison with the results of Heimsath et al. (2012).

10   Figure 3 illustrates the effects of topography on tectonic stresses only. Gravitational stresses can be included in the model by superposing the analytic solutions of Savage and Swolfs (1986) (their equations (34) and (35)) with the solutions of Savage et al. (1985) for the effects of topography on gravitational stresses (their equations (39) and (40)). The result is a three-dimensional phase space of solutions corresponding to different values of the regional tectonic stress N_1, the characteristic gravitational stress ρgb (where ρ is the density of rock, g is the acceleration due to gravity, and b is the ridge

15 height), and the Poisson ratio μ. The effects of including gravitational stresses are (1) to increase the compression at depth via the lithostatic term (at soil depths this corresponds to an addition of ~10 kPa, which is negligible compared to the regional compressive stress of ~10 MPa in the SGM), and 2) to increase the compressive stresses near the point of inflection on hillslopes (e.g., Fig. 2a of Savage et al., 1985). These modifications do not alter the first-order behavior illustrated in Figure 3 for rocks close to the surface that are not close to hollows or other points of inflection. Section 3 provides additional

20 discussion of the assumptions and alternative approaches to modeling topographically induced stresses.

        The fit of the solid curve in Figure 2A to ~~P₀,resid~~P_r values is based on equation (3), together with an assumption that the transition from compressive to tensile stresses triggers ~~an~~a step increase in ~~P₀,resid~~P_r values over a small range of S_{av} values in the vicinity of the transition from compression to tension:

6

$$P_{0,S} = \begin{cases} P_{0,t}\left(1 - \frac{\sigma_{xx}}{N_t}\right) & \text{if } S_{av} \leq S_t \\ P_{0,h}\left(1 - \frac{\sigma_{xx}}{N_t}\right) & \text{if } S_{av} > S_h \\ \left(P_{0,t} + \left(P_{0,h} - P_{0,t}\right)\frac{S_{av}-S_t}{S_h-S_t}\right)\left(1 - \frac{\sigma_{xx}}{N_t}\right) & \text{if } S_t \leq S_{av} < S_h \end{cases} \qquad (4)$$

$$P_{r,S} = \begin{cases} P_{r,l}\left(1 - \frac{\sigma_{xx}}{N_1}\right) & \text{if } S_{av} \leq S_l \\ P_{r,h}\left(1 - \frac{\sigma_{xx}}{N_1}\right) & \text{if } S_{av} > S_h \\ \left(P_{r,l} + \left(P_{r,h} - P_{r,l}\right)\frac{S_{av}-S_l}{S_h-S_l}\right)\left(1 - \frac{\sigma_{xx}}{N_1}\right) & \text{if } S_l \leq S_{av} < S_h \end{cases} \qquad (4)$$

where $P_{r,S}$ denotes the model for the dependence of P_r values on S_{av}, $P_{r,l}$ and $P_{r,h}$ are coefficients defining the low and high values of P_r, and S_l and S_h are the average slopes defining the range over which P_r values increase from low to high values across the transition from compression to tension. $P_{r,l}$ and $P_{r,h}$ were determined to be 170 m/Myr and 500 m/Myr based on least-squares minimization to the data (data from elevations above 2300 m were excluded because of the climatic influence described below). S_l and S_h were chosen to be 30° and 32°, respectively, to characterize the abrupt increase in P_r values in the vicinity of 30°.

In addition to the average slope control associated with the topographically induced stress fracture opening process, a climatic control on P_r values can be identified using cluster analysis. This type of analysis involves identifying clusters in the data defined by distinctive values of the independent variables that also have different mean values of the dependent variable. The four points colored in blue in Figure 2A are the four highest elevation samples in the dataset, with elevations $\geq$ 2300 m a.s.l. The logarithms (base 10) of this cluster have a mean value of -0.40 after subtracting the logarithms of $P_{r,S}$ to account for the average slope control on P_{resid} values, compared with a mean of 0.00 for the logarithms of the remaining data points with $S_{av} > 30°$ (also with the logarithms of $P_{r,S}$ subtracted). Assuming a significance level of 0.05, the null hypothesis that the cluster of blue points has a mean that is indistinguishable from that of the remaining points with $S_{av} > 30°$ can be rejected based on the standard t test with unequal variances ($t = 0.021$).

Figures 4A-4C illustrate the mean annual temperature (MAT), mean annual precipitation (MAP), and existing vegetation height (EVH) for the central portion of the SGM. Above elevations of approximately 1800 m a.s.l., vegetation height decreases systematically with increasing elevation (Fig. 4D). This limitation is likely to be primarily a result of temperature limitations on vegetation growth because MAP increases with elevation up to and including the highest

elevations of the range. This result is consistent with the hypothesis that vegetation is a key driver of soil production. The decrease in P_0 values with elevation is likely to be gradual rather than abrupt, and indeed there is evidence of a peak in the climatic control of P_0 values. Figure 4E plots the ratio of ~~$P_{0,resid}$~~P_r to ~~P_0~~$P_{r,\text{S}}$ as a function of elevation. The closed circles are binned averages of the data (each bin equals 100 m in elevation). The ratio of ~~$P_{0,resid}$~~P_r to ~~P_0~~$P_{r,\text{S}}$ (equivalent to the residuals

5   under log transformation after the effects of average slope are removed) increases, on average, and then decreases within the range of elevations between 1500 and 2600 m, broadly similar to the trend of EVH (Fig. 4D).

Local variability in ~~P_0~~P_r estimates due to variations in soil thickness, mineralogical variations within a given lithology, spatial variations in fracture density, etc. can be minimized by averaging ~~P_0~~P_r values (not including the four highest-elevation points because of the climatic control) from locations that have the same average slope (Fig. 2C). This

10   process tends to average data from the same local cluster since local clusters often have average slopes that are both equal within the cluster and different from other clusters. Figure 2C demonstrates that the predictions of the topographically induced stress fracture opening hypothesis are consistent with the observed dependence of ~~$P_{0,resid}$~~P_r values on S_{av} values.

The average slope and climatic controls on ~~P_0~~P_r values can be combined into a single predictive equation for ~~P_0~~P_r values:

$$\sout{P_{0,\text{pred}}}P_{r,\text{pred}} = \sout{P_{0,s}}P_{r,s}\,C \tag{5}$$

where ~~P_0~~$P_{r,\text{pred}}$ denotes predicted values for ~~P_0~~P_r, C is a climatic index defined as 1 for $z < 2300$ m and 0.4 (i.e., the ratio of the mean of the logarithms of the data for $z > 2300$ m to the mean of the logarithms of remaining data points with $S_{av} > 30°$) for $z > 2300$ m. A regression of ~~P_0~~$P_{r,\text{pred}}$ values to ~~$P_{0,resid}$~~P_r values yields an R^2 of 0.50 (Fig. 2D). When data with equal S_{av}

20   values are averaged (i.e., the filled circles in Fig. 2D), the resulting R^2 value is 0.87.

The results of this section demonstrate that average slope and climate exert controls on P_0 values in the SGM. Although I did not find additional controls that were clearly distinct from these, it is worth discussing additional controls that I tested for. The data points colored in gray in Figure 2B are from the three rock types most resistant to weathering as determined by Spotila et al. (2002): granite, anorthosite, and the Mount Lowe intrusive suite. Spotila et al. (2002) also

25   identified gabbro as a relatively resistant rock in the SGM, but no soil production rates are available from this rock type.

Figure 2B suggests that lithology might exert some control on $P_0 P_r$ values. Specifically, 7 samples from the more resistant lithologies sit above the least-squares fit of equation (4) to the data, while 13 (including the 7 lowest P_0 values) sit below the least-squares fit. However, the null hypothesis that the residuals of the gray cluster after the effects of average slope are removed has a mean that is indistinguishable from the residuals of the remaining points (colored black in Figure 2B) cannot

5   be rejected ($t = 0.21$).

Many studies have proposed a relationship between fracture density and bedrock weatherability on the basis that fractures provide additional surface area for chemical weathering and pathways for physical weathering agents to penetrate into the bedrock or intact regolith (e.g., Molnar, 2004; Molnar et al., 2007; Goodfellow et al., 2014; Roy et al., 2016a,b). The difference in erosion rates between the SGM and adjacent San Bernadino Mountains, for example, has been attributed in part

10   to differences in fracture density between these ranges (Lifton and Chase, 1992; Spotila et al., 2002). As such, it is reasonable to hypothesize that differences in $P_0 P_r$ values might result from spatial variations in fracture density within each range. I computed a bedrock damage index D based on the concept that $P_0 P_r$ values increase in bedrock that is more pervasively fractured, together with the fact that bedrock fracture densities are correlated with local fault density in the SGM (Chester et al., 2005; Savage and Brodsky, 2011). Savage and Brodsky (2011) documented that bedrock fracture density

15   decreases as a power-law function of distance from small isolated faults, i.e. as $r^{-0.8}$ where r is the distance from the fault. Fracture densities around larger faults and faults surrounded by secondary fault networks can be modeled as a superposition of $r^{-0.8}$ decays from all fault strands (Savage and Brodsky, 2011). Chester et al. (2005) documented similar power-law relationships between bedrock fracture density and local fault density in the SGM specifically. I define the bedrock damage index D (Fig. 5A) as the sum of the inverse distances, raised to an exponent 0.8, from the point where the D value is being

20   computed to every pixel in the study area were a fault is located:

$$D = \sum_{\mathbf{x}'} \left(\Delta x / |\mathbf{x} - \mathbf{x}'| \right)^{0.8} \qquad\qquad (6)$$

where Δx is the pixel width, $\mathbf{x}$ is the map location where bedrock damage is being computed, and $\mathbf{x}'$ is the location of each mapped pixel in SGM where a fault exists. D has units of length since it is the sum of all fault lengths in the vicinity of a point, weighted by a power function of inverse distance. Equation (6) honors the roles of both the distance to and the local

25   density of faults documented by Savage and Brodsky (2011) because longer faults and/or more mature fault zones with many

secondary faults have more pixels that contribute to the summation. The fact that a relationship exists between $P_{0,resid}P_r$ values and D (Fig. 5B, $p = 0.035$) and between D and S_{av} (Fig. 5C, $p = 0.015$) suggests that some of the control by average slope that I have attributed to the topographically induced stress fracture opening process may reflect differences in the density of pre-existing fractures related to local fault density. However, the much higher R^2 value of the relationship between

5 $P_{0,resid}P_r$ and $P_0P_{r,pred}$ ($R^2 = 0.50$) compared to that for the relationship between $P_{0,resid}P_r$ and D ($R^2 = 0.08$) suggests that the topographically induced stress fracture opening process is the dominant mechanism controlling P_0P_r values in the SGM. In addition, this process has a stronger theoretical foundation.

2.2 Relating potential soil production rates to erosion rates and topographic steepness in the SGM

10   In this section I invoke a balance between soil production and transport at the hillslope scale in order to illustrate the interrelationships among potential soil production rates, erosion rates, soil thicknesses, and average slopes across the SGM. The conceptual model explored in this section is based on the hypothesis that the average slope depends on the long-term difference between uplift and erosion rates. Uplift rates (assumed here to be equal to exhumation rates) are lower in the western portion of the SGM and higher in the eastern portion (Spotila et al., 2002, their Fig. 7b). As average slope increases

15 in areas with higher uplift rates, erosion rates increase and soils become thinner. Both of these responses represent negative feedback mechanisms that tend to decrease the differences that would otherwise exist between uplift and erosion rates and between erosion rates and soil production rates. If the uplift rate exceeds the potential soil production rate, soil thickness becomes zero and soil production and erosion rates can no longer increase with increasing slope (in the absence of widespread landsliding in bedrock or intact regolith). In such cases, topography with cliffs or steps may form (e.g.,

20 Wahrhaftig, 1965; Strudley et al., 2006; Jessup et al., 2010). However, if the potential soil production rate increases with average slope via the topographically induced stress fracture opening process, the transition to bare landscapes can be delayed or prevented as Heimsath et al. (2012) proposed. This represents an additional negative feedback or adjustment mechanism. At the highest elevations of the range, soil production is slower, most likely due to temperature limitations on vegetation growth. The interrelationship between these variables can be quantified without explicit knowledge of the uplift

25 rate, since the relationship between soil thickness and average slope implicitly accounts for the uplift rate (i.e., a smaller

difference between uplift and erosion rates is characterized by a thinner soil). This conceptual model predicts positive correlations among potential soil production rates, erosion rates, and topographic steepness, and negative correlations of all of these variables with soil thickness.

Equation (5), in combination with modified versions of equations (9)&(11) of Pelletier and Rasmussen (2009), i.e.,

5
$$\cancel{P_0 e^{-h/h_0} = E} \; P_r e^{-h/h_0} = E \tag{7}$$

and

$$\frac{\kappa S_{av}}{1 - (S_{av}/S_c)^2} = EL, \tag{8}$$

predict spatial variations in erosion rates and topographic steepness associated with spatial variations in $\cancel{P_0} P_r$ values. In

10 equations (7)&(8), κ is a sediment transport coefficient (m^2/Myr) and L is a mean hillslope length (m). Equation (8) assumes a steady state balance between soil production and erosion, modeled via the nonlinear slope-dependent sediment flux model of Roering et al. (1999) at the hillslope scale. Equation (8) assumes that the mean slope gradient at the base of hillslopes (where the sediment flux leaves the slope) of a given area can be approximated by the average slope.

Spatial variations in erosion rates can be estimated using $\cancel{P_0} P_r$ values predicted by equation (5) if spatial variations

15 in soil thickness can also be estimated. To do this, I developed an empirical relationship between soil thickness and slope gradient derived from the Heimsath et al. (2012) dataset (Fig. 6):

$$h = \frac{h_1}{S_{av}^b}, \tag{9}$$

with best-fit coefficients of $b = 1.0$ and $h_1 = 0.06$ m ($R^2 = 0.18$, $p = 0.001$). For this regression, I shifted the soil thickness in areas with no soil upward to a small finite value (0.03 m). These areas have no soil today, but must have had some soil over

20 geologic time scales or else no erosion would occur. Also, without some shift, the 10 data points with $h = 0$ cannot be used, biasing the analysis towards areas that have soil cover today. The 0.03 m value was chosen because this is the minimum finite soil thickness measured by Heimsath et al. (2012).

Using equation (9) as a substitution, equations (7)&(8) can be combined to obtain a single equation for topographic steepness, S_{av}:

$$\frac{S_{\text{av}}}{1-(S_{\text{av}}/S_{\text{c}})^2} = \frac{L}{\kappa}\overline{P_{\text{0,pred}}}P_{\text{r,pred}}\exp\left(-\frac{h_1}{h_0 S_{av}^b}\right)$$

(10)

Given a map of steepness obtained by solving equation (10), soil thicknesses and erosion rates can be mapped using equations (7) and (8), respectively. Note that the S_{av} value obtained by solving equation (10) is not a prediction in the usual sense, since S_{av} is an input to eqn. (10) via $\overline{P_0}\underline{P_{\text{r}}}_{\text{,pred}}$. The model can be considered to capture the effects of topographic steepness if the predicted and observed values of S_{av} have broadly similar absolute values and patterns of spatial variation.

Equations (7)&(8) are the same as equations (9)&(11) of Pelletier and Rasmussen (2009) except that their equation (9) included a term representing the bedrock-soil density contrast related to a slightly different definition of $\overline{P_0}\underline{P_{\text{r}}}$ (termed P_0 in Pelletier and Rasmussen (2009)) and their equation (11) assumed a depth- and slope-dependent transport relation. Here I use a slope-dependent relation because depth-dependent models depend on the average soil depth *when soil is present* (because soil must be present for transport to occur), which cannot be determined for locations where soil thickness is currently zero.

The S_{av} values predicted by equation (10) (Fig. 7C) reproduce the observed first-order patterns of topographic steepness (Fig. 7C) if L/κ = 0.005 Myr/m and S_{c} = 0.8 are used. The value S_{c} = 0.8 was chosen because it is in the middle of the range of values (i.e., 0.78-0.83) that Grieve et al. (2016) obtained for steep landscapes in California and Oregon. With this value for S_{c}, the best-fit value for L/κ was determined by minimizing the least-squares error between the model prediction (Fig. 7B) and observed variations in average slope (Fig. 7C). Predicted and measured S_{av} values are lowest in the Western block and higher in the Sierra Madre, Tujunga, and Baldy blocks. Soil thicknesses predicted by the model correlate inversely with slopes and P_0 values (Fig. 7D). Erosion rates (Fig. 7E) closely follow $\overline{P_0}\underline{P_{\text{r}}}$ values, but are lower in absolute value, reflecting the buffering effect of soil on bedrock physical weathering processes.

3 Discussion

This paper adopts a stepwise regression and cluster analysis approach that builds upon the regression analysis that Heimsath et al. (2012) used to characterize the dependence of soil production rate on soil thickness. Stepwise regression is a standard approach in statistics in which the residuals of a statistical regression are computed and additional controls tested

for. I did not apply simultaneous multivariate linear regression (with or without log transformation) because such an approach would have been inconsistent with the complex nonlinear relationships in the data documented by Heimsath et al. (2012) and the analyses presented here.

Estimating P_r values using the residuals of the regressions of Heimsath et al. (2012) assumes that h_0 has sufficiently limited variation within the two subsets of the study site considered by Heimsath et al. (2012) (i.e., those with S_{av} values above and below 30°) that any such variation would not affect the conclusions of the paper. For example, in order for the relationship between P_r and S_{av} (i.e., Figs. 2A-2C) to be significantly affected by variations in h_0, h_0 would have to have a systematic dependence on S_{av}. For example, if systematically lower values of h_0 occur at steeper slopes and this effect is not accounted for, the result could be a biasing of P_r values downward in such regions. Heimsath et al. (2012) clearly demonstrated that no such systematic dependence exists. These authors considered two end member slope regimes and found that the average h_0 values for these two regions differed by only 0.05 m (0.32 m vs. 0.37 m). At a soil thickness of 0.3 m, this difference corresponds to P_r differences of approximately 10% (i.e., exp(-0.30/0.32) vs. exp(-0.30/0.37)). This difference is more than 100 times smaller than the variation in P_r values. The difference becomes even smaller for soils thinner than 0.3 m.

The effect of topographically induced stresses on regolith production is a rapidly evolving field at the boundaries among geomorphology, geophysics, and structural geology. The results presented here, based on the Savage and Swolfs (1986) model, represents just one possible approach to the problem. Miller and Dunne (1998), for example, modified the Savage and Swolfs (1986) solutions to account for cases with vertical compressive stress gradients (their parameter k) larger than 1. Data from the SGM and the adjacent southwestern Mojave Desert indicate that the vertical gradient of horizontal stress in the SGM is likely less than one. Sbar et al. (1979) measured mean maximum compressive stresses at the surface equal to 16 MPa, which is similar to values measured at depths of 100-200 m obtained by Zoback et al. (1980) (their Figs. 7&10). As such, the Savage and Swolfs (1986) approach is likely to be appropriate for the SGM. In addition to the effects of variations in the depth gradient of stress, fractures can open beneath hillslopes in a direction perpendicular to the slope, parallel to the slope, or in shear. The criteria for each of these strains depends on different components and/or derivatives of the stress field. For example, Martel (2006, 2011) emphasized the vertical gradient of vertical stress, which depends on the

topographic curvature instead of the slope, in driving fracturing parallel to the surface, while St. Clair et al. (2015) emphasized the ratio of the horizontal stress to the spacing between ridges and valleys. More research is needed in the SGM and elsewhere to better understand the response of bedrock and intact regolith to the 3D stress field. However, all studies agree that the extent of one or more fracture opening modes increases with topographic slope and/or curvature, often with a

5   threshold change from compression to tension above a critical value of topographic ruggedness.

The results presented here provide a process-based understanding of the dependence of potential soil production rates on topographic steepness documented by Heimsath et al. (2012) in the SGM. These authors proposed a negative feedback in which high erosion rates trigger higher potential soil production rates, with the result that soil cover may more persistent than previously thought. The results presented here suggest that, in the SGM, the release of compressive stress in

10   steep landscapes causes fractures beneath ridges to open, thereby allowing weathering agents to penetrate into the bedrock or intact regolith more readily. The fact that this process requires a regional compressive stress state suggests that this it is not likely to be equally important everywhere on Earth. In cases of low regional compression or extension, the development of rugged topography in rocks with pre-existing fractures is not likely to be significant in promoting fracture opening in the rocks beneath hillslopes.

15   Heimsath et al. (2012) argued that ~~P_0~~P_r values (analogous to what they termed SPR_{max} values) increase with erosion rates not just in the SGM, but globally based on the strong correlation between P and E values (their Fig. 4b). However, the results of this paper suggest that ~~the~~ it is slope that controls P_r values, not erosion rate. Slope and erosion rate are highly correlated in the SGM, but this correlation is not universal. The results of this paper also suggest that the process ~~that~~ by which slope leads to an increase in P_0 values ~~with increasing topographic ruggedness in the SGM~~ in the SGM (i.e.,

20   topographically induced stress opening of fractures) is likely not operative ~~everywhere~~ in extensional or neutral-stress settings. As such, other factors might explain the global correlation between P and E values. For example, erosion rates may be limited by ~~P_0~~P_r values (since erosion cannot occur faster than soil is produced in the absence of widespread landsliding in bedrock or intact regolith). ~~P_0~~P_r values are a function of climate, with values exceeding 1000 m/Myr in humid climates (Pelletier and Rasmussen, 2009; Larsen et al., 2014). As such, the global correlation between P and E values may, in part, be

25   a result of water availability being important for both soil production and erosion processes. If erosion rates cannot keep pace

with erosion rates, stepped topography can and does form in some cases (e.g., Wahrhaftig, 1965; Strudley et al., 2006; Jessup et al., 2010), leading to a reduction in erosion rates (as evidenced by lower soil production rates in bare areas relative to soil-covered areas (Hahm et al., 2014)) despite locally steeper slopes. In such cases, P and E values are still correlated because erosion cannot occur at rates higher than $P_0 P_r$.

4 Conclusions

In this paper I estimated spatial variations in the potential soil production rate, $P_0 P_r$, using cosmogenic-radionuclide-derived soil production rates from the central San Gabriel Mountains of California published by Heimsath et al. (2012). The results demonstrate that trends in the data are consistent with the hypothesis that topographically induced stresses cause pre-existing fractures to open beneath steeper hillslopes. This model predicts an abrupt increase in $P_0 P_r$ values close to the average slope (approximately 30°) where an increase is observed in the data. After the effects of topographically induced stress are accounted for, a limitation on $P_0 P_r$ values can be detected at the highest elevations of the range, where vegetation growth is limited by temperature. There is some evidence that lithology and local fault density may also influence potential soil production rates, but the null hypotheses that these processes are not significant cannot be ruled out with given a threshold statistical significance (false positive rate) of 0.05, or they cannot be clearly distinguished from other controls. The results of this paper demonstrate that $P_0 P_r$ values are solely dependent on climate and rock characteristics, but that rock characteristics evolve with topographic ruggedness in compressive stress environments. These results provide a useful foundation for additional targeted cosmogenic-radionuclide analyses in the San Gabriel Mountains and for the incorporation of methods that can further test the topographically induced stress fracture opening hypothesis such as shallow seismic refraction surveys and 3D stress modeling.

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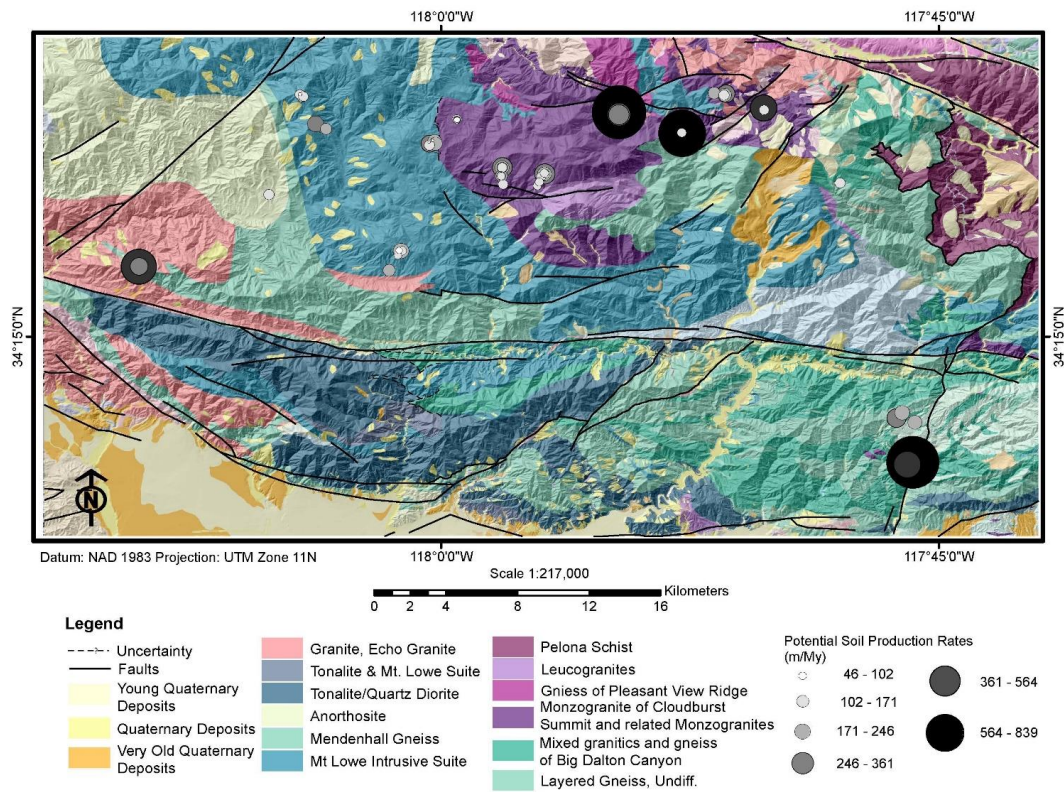
Figure 1. Geologic map of the central San Gabriel Mountains, California. Potential soil production rates inferred from the data of Heimsath et al. (2012) are also shown. Lithologic units were compiled using Yerkes and Campbell (2005), Morton and Miller (2003), and Figure 3 of Nourse (2002). Faults were mapped from Morton and Miller (2003) and the Quaternary fault and fold database of the United States (U.S. Geological Survey and California Geological Survey, 2006).

A

$b/a = 0, S_{av} = 0$

$b/a = 1, S_{av} = 14°$

$b/a = 2, S_{av} = 27°$

$b/a = 2.5, S_{av} = 32°$

σ_{xx}/N_1

-1   -0.25 -0.05 0 0.05 0.25   1

B

compression

tension

σ_{xx}/N_1

S_{av} (°)

Figure 3. **Plots of** $P_{0,resid}$ and their relationship to average slope, S_{av}, and other potential controlling factors. (A) Plot of $P_{0,resid}$ P_r values versus S_{av}. Data points colored blue are from the highest elevations of the range ($z > 2300$ m). (B) The same plot as (A), except that data points are colored according to whether they from rocks that are relatively more resistant (gray) or less resistant (black) to weathering. (C) Plot of $P_{0,resid}$ P_r values averaged for each value of S_{av}. In (A) and (B), error bars represent the uncertainty of each data point, while in (C) the error bar represents the standard deviation of the data points averaged for each S_{av} value. (D) Plot of $P_{0,resid}$ P_r versus values predicted from equation (5). Unfilled circles show individual data points, while filled circles represent the averaged data plotted in (C).

Figure 3.

Figure 4. Climate and vegetation cover of the central San Gabriel Mountains. Color maps of (A) mean annual temperature (MAT) and (B) mean annual precipitation (MAP) from the PRISM dataset (Daly et al., 2001). (C) Color map of mean existing vegetation height (EVH) from the U.S. Geological Survey LANDFIRE database (U.S.G.S., 2016). (D) Plot of mean EVH versus elevation above sea level, z, using the data illustrated in (C). (E) Plot of the ratio of $P_{0,resid}P_r$ to $P_0 P_{r,S}$ as a function of elevation. Filled circles are binned averages of the data (each bin equals 100 m in elevation).
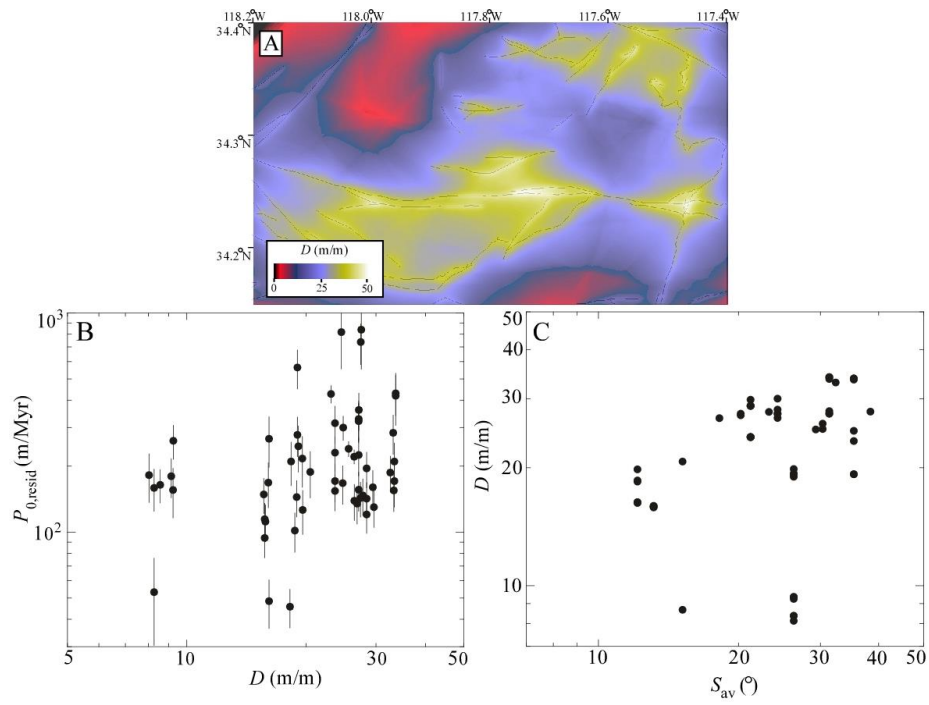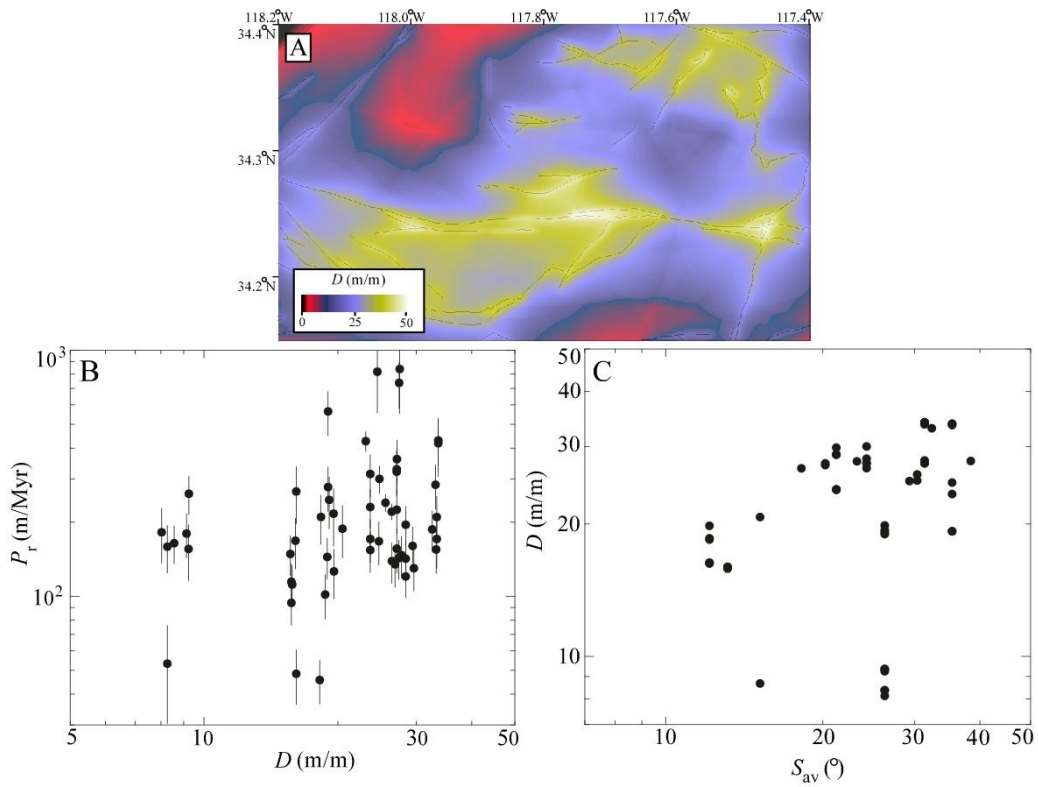
Figure 5. Map of the bedrock damage index, *D*, and its correlation with *S*ₐᵥ. (A) Color map of spatial variations *D*. (B) Plot of *D* versus *S*ₐᵥ for the 57 sample locations of Heimsath et al. (2012).
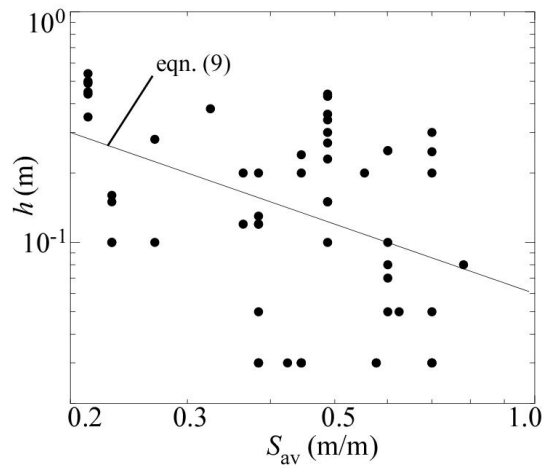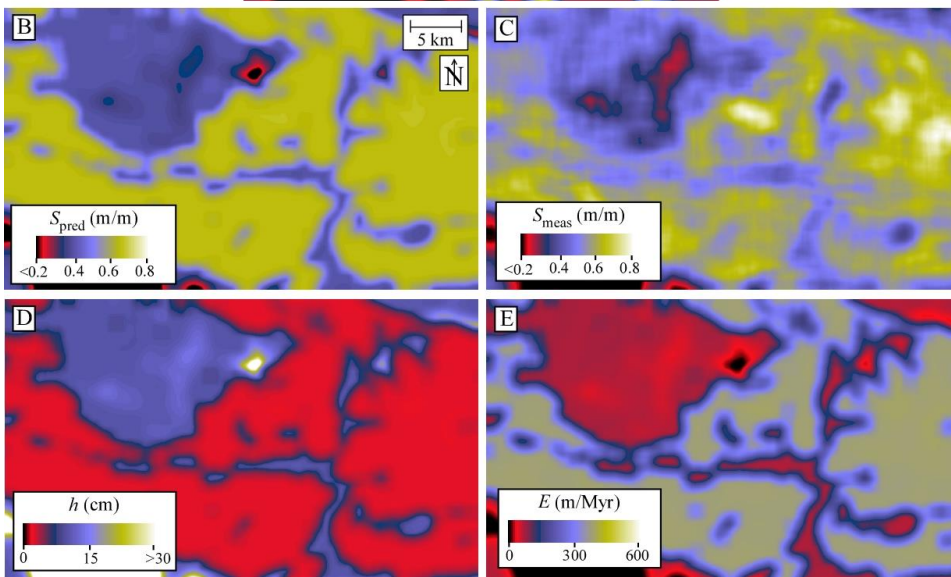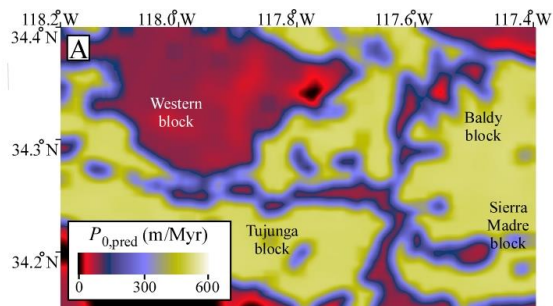
Figure 6. Plot of soil thickness, h, as a function of average slope, S_{av}. The least-squares power-law fit to the data (equation (9)) is also shown.
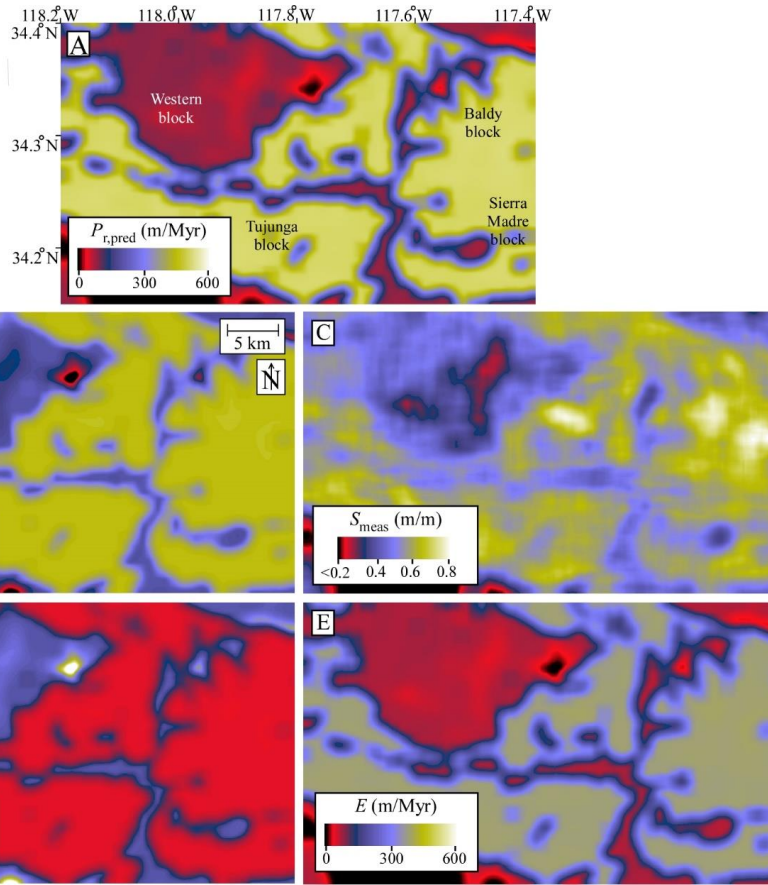
Figure 7. Color maps illustrating the predicted potential soil production rate from equation (5) ($P_0 P_{r,pred}$), predicted and observed values of average slope, S_{av}, soil thickness, h, and erosion rate, E. (A) Color map of $P_0 P_{r,pred}$ values estimated from equation (5). (B) Color map of S_{av} values predicted by equation (10), smoothed by a moving average filter with a 1-km length scale to emphasize patterns at the landscape scale. (C) Color map of actual (DEM-derived) S_{av} values, smoothed in the same manner as (B). (D) Color map of soil thicknesses, h, predicted by equation (9). (E) Color map of erosion rates, E predicted by equation (7).