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Notes

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ABSTRACT

Alluvial riverbed elevation responds to the balance between sediment supply and transport capacity, which is largely dependent on climate and its translation into fluvial discharge. We examine these relations using U.S. Geological Survey streamflow and channel measurements in conjunction with basin characteristics for 915 reference ("least disturbed") measurement stations across the conterminous United States for the period A.D. 1950–2011. We find that (1) 68% of stations have bed elevation change (*BEC*) trends ($p < 0.05$) with median values of +0.5 cm/yr for aggradation and –0.6 cm/yr for degradation, with no obvious relation to drainage basin structure, physiography, or lithology; (2) *BEC* correlates with drainage basin area; (3) high-flow variability (Q_{90}/Q_{50}, where Q is discharge and 90 and 50 are annual flow percentiles) translates directly into the magnitude, though not the direction, of *BEC*, after accounting for the scale dependence; (4) Q_{90}/Q_{50} declines systematically from dry to wet climates, producing disproportionately high rates of *BEC* in drier regions; and (5) marked increases in precipitation and streamflow occurred disproportionately at dry sites, while streamflow declined disproportionately at wet sites. Climatic shifts in streamflow have the potential to increase/decrease sediment flux and thus affect riverbed elevation by altering flood frequency. These unforeseen responses of bed elevation to climate and climate change have important implications for sediment budgets, longitudinal profiles, ecology, and river management.

INTRODUCTION

Riverbed elevation is a fundamental boundary condition for landscape evolution (Tucker and Slingerland, 1997), alluvial stratigraphy (Daniels, 2008), flood risk, and aquatic habitat (ASCE Task Committee on Sediment Transport and Aquatic Habitats, Sedimentation Committee, 1992). Over societally relevant time scales, anthropogenic activity (Syvitski et al., 2005) and climate (Inman and Jenkins, 1999) govern changes in sediment production and flux through fluvial systems, and by extension, they affect alluvial riverbeds (Daniels, 2008; Simon, 1989; Williams and Wolman, 1984). But in contrast to substantial work documenting anthropogenic impacts in alluvial rivers (Gregory, 2006), we have no corresponding understanding of climatic riverbed signatures, due to a lack of systematic observations spanning gradients of drainage area and climate (Kochel, 1988).

Previous work has suggested that climate etches its signature in river channels (Hartshorn et al., 2002; Molnar et al., 2006; Stark et al., 2010), but it is not clear over what time scales this imprint develops, or how precipitation changes are reflected in alluvial riverbed elevation. Precipitation generates runoff and erodes sediment from various parts of a drainage basin; a proportion of it forms the riverbed, with grain sizes that provide resistance to subsequent transport. This bed sediment is transported through the river network by streamflow once it exceeds the entrainment threshold. The frequency of such transport is affected by climatic regimes (Groisman et al., 2001; Molnar et al., 2006; O'Connor and Costa, 2004; Pitlick, 1994), producing reach-scale scour/fill (short-term variability) or aggradation/degradation (long-term trends) over multiple storm cycles. For steady basin conditions, changes in riverbed elevation reflect the balance between supply and transport capacity, i.e., net sediment flux through the fluvial system.

Over longer time scales (10^4–10^6 yr), net sediment flux and fluvial erosion have been generalized using landscape-evolution models to assess climate impacts on river incision and drainage network development (e.g., Tucker and Slingerland, 1997). These models provide insight on the long-term impacts of increasing humidity/aridity on channels, but are not appropriate for understanding the short-term responses of alluvial riverbeds to changes in climate, which are transient over their time domains. Other research has correlated modern records of precipitation (*P*) or discharge (*Q*) with vertical (Hartshorn et al., 2002) and lateral (Stark et al., 2010) erosion rates in bedrock rivers, but the spatial and temporal mismatches between *P*, *Q*, and erosion data sets make it difficult to interpret transient responses of river systems to climate. Likewise, concurrent paleorecords of climate and sedimentation (10^3–10^5 yr) have been employed to provide inferences of past landscape response (Knox, 2000). However, such analyses are complicated by temporal biases in deposited sediments (Jerolmack and Paola, 2010; Sadler, 1981), poorly constrained prehistoric climate records, and feedbacks between riverbed elevation and sedimentation rates (Daniels, 2008).

In the United States there is excellent documentation of changes in the magnitude and variability of *P* (Karl et al., 1995), alteration of its timing and form (Knowles et al., 2006), and trends in the frequency of extreme rainfall and associated floods (Groisman et al., 2001) for the past several decades. Yet despite decades of *P* and *Q* measurement, the links between climate and contemporaneous changes in river morphology remain unexplored at broad spatial scales. This shortcoming thwarts short-term prediction of how climate change might affect flood risk (e.g., ratios of flood stage to floodplain elevation), longitudinal sediment storage, and the spatial distribution and stability of aquatic habitat (e.g., frequency of bed disturbance events and amount of in-channel sediment storage), as well as long-term understanding of drainage basin development (e.g., river longitudinal profile shape).

In this paper, we examine relationships between riverbed elevation time series, basin characteristics, climate, and climate change. Our purpose is to (1) investigate broad-scale controls on riverbed elevation trends; (2) assess the variability in this signature across climatic gradients; and (3) determine how these trends may be affected by climate changes.

METHODS

We analyzed publicly available U.S. Geological Survey streamflow measurements and daily flow data for reference (i.e., a classification that denotes watersheds that are minimally disturbed by human influences) gaging stations in the Gages-II data set (Falcone, 2011) for 1950–2011. Mean channel bed elevation (*BE*) was calculated from each *Q* measurement as: $BE = h - (A/w)$, where h is flow stage above local datum, A is flow area, and w is flow width, assuming a rectangular cross section. The values for h, A, and w are provided for each measurement in the database and we only utilized measurements from stable gage sites since 1950. Trends were investigated by evaluating these calculated *BE* values for individual stations over time. For gaging sites with time series trends (bed elevation trend, *BET*) in *BE* ($p < 0.05$), we derived annualized rates of aggradation/degradation. For those without significant trends, we calculated the interquartile range of *BE* (bed elevation variability, *BEV*). We also computed annual *Q* percentiles

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(Q_{90} and Q_{50} in m^3/s) from daily streamflow data and temporal trends ($p < 0.05$) in these percentiles (Q_{90}trend and Q_{50}trend in $m^3/s/yr$).

We assessed *BET* and *BEV* with respect to Q_{90}, Q_{50}, Q_{90}trend, Q_{50}trend, a measure of high-flow variability (Q_{90}/Q_{50}), mean annual precipitation (mean*P* in mm), annualized trend ($p < 0.05$) in mean precipitation (*P*trend in mm/yr) (Mitchell and Jones, 2005), as well as drainage basin characteristics that are generally assumed to control river incision, including drainage area (*DA* in km^2), site elevation (m above sea level, masl), mean watershed slope (%), drainage density (km/km^2), and dominant basin lithology (see the GSA Data Repository[1]).

RESULTS AND DISCUSSION

We find decadal trends of bed elevation change for 68% of the 915 stations analyzed. Of these, riverbeds were aggrading (*BET*+) at 38% and degrading (*BET*–) at 62%. Figures 1A and 1B show a broad mix of *BET* and *BEV* sites, respectively, with no obvious systematic geographic biases. In addition, there are regions where *BET*+ and *BET*– sites are colocated, as well as local concentrations of aggrading (Pacific Northwest) and degrading (Appalachian Mountains) stations within similar mean*P* regimes, but these also occur in regions that have had contrasting trends in mean*P* (Fig. 1B; Groisman et al., 2001).

BE time series reveal that high-frequency fluctuations in riverbed level are typically caused by short-term variability in local hydrology (expressed as *BEV* at otherwise stable sites; Fig. 1B). Over decadal time scales, however, *BE* trends capture systemic shifts in climate and/or sediment supply (e.g., decreasing *BE*, concomitant with an upward shift in Q_{50} and Q_{90}; Fig. 1C). These factors suggest that changes in bed elevation may be intrinsically linked to spatial and temporal differences in the frequency of sediment-transporting events expressed between climatic regions (Baker, 1977; Knowles et al., 2006; Molnar et al., 2006; Tucker and Bras, 2000).

Drainage Basin Controls

Models of long-term landscape evolution often assume constant basin-wide sediment supply, so stream power (e.g., *Q**channel gradient) controls rates of bed elevation change (*BEC*) and further, *Q* scales with *DA*. We find that none of the following variables correlates with the magnitude of *BET* or *BEV*: mean watershed slope, drainage density, relief, elevation, mean*P*, or lithology (Fig. DR1 in the Data Repository). However, our results show that

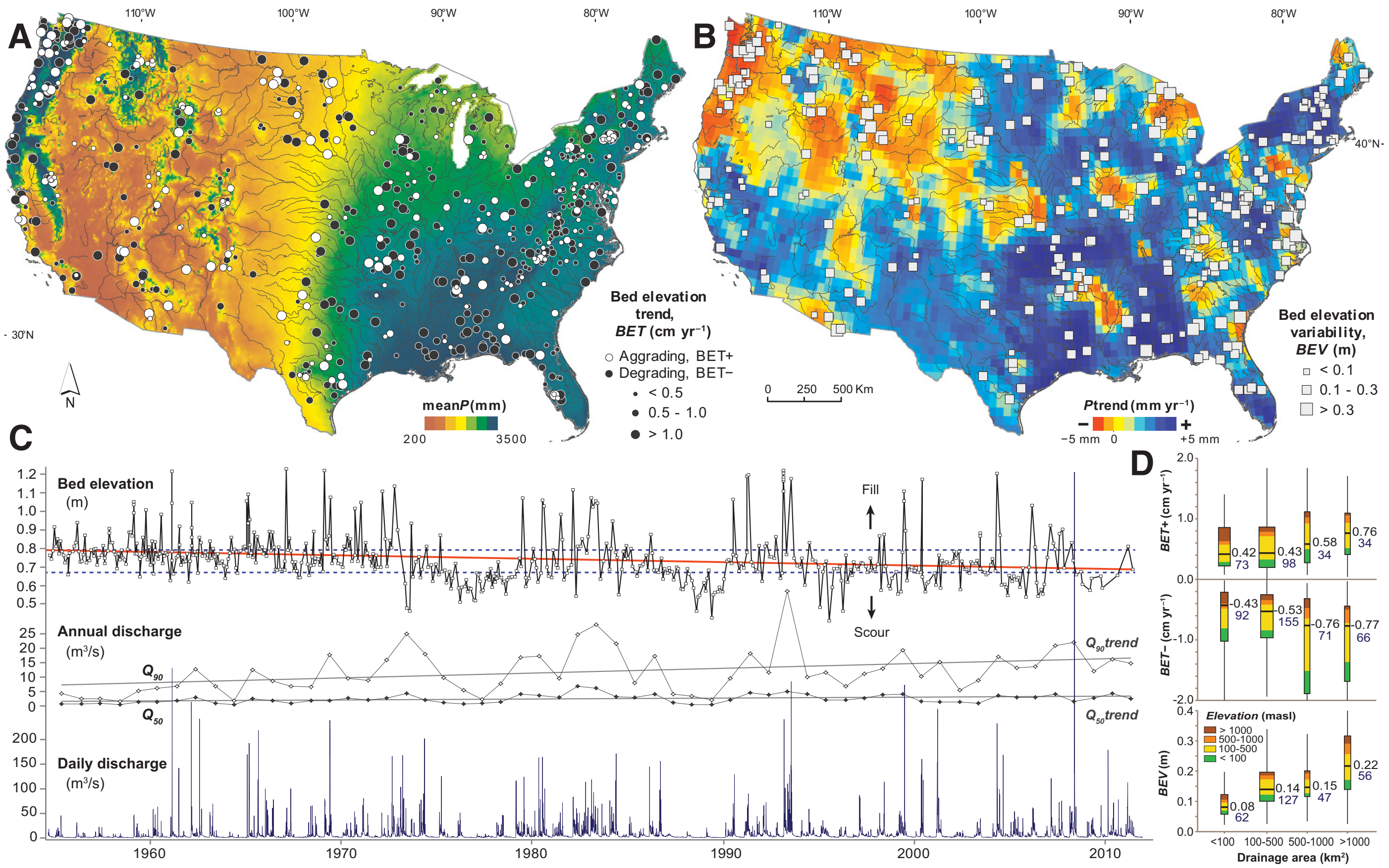

Figure 1. A: Map of bed elevation trend (*BET*) against background of mean annual precipitation (mean*P*) values (see the Data Repository [see footnote 1]). B: Map of bed elevation variability (*BEV*) against background of annualized trend in mean precipitation (*P*trend) values. C: Time series at site 05458000: daily discharge (*Q*, m^3/s), annual *Q* percentiles (Q_{50} and Q_{90}, m^3/s), and *BE* (m) (for statistics, see the Data Repository). Dashed lines represent interquartile range of *BE*, used in calculation of *BEV*, and red line marks degrading *BE* trend. D: Boxplots of aggradation (*BET+*), degradation (*BET*–), and *BEV* versus drainage area. Boxes, depicting distributional interquartile range, are variable width based on number of stations in each bin. Color stripes indicate proportionality of sites within each category. Values to right of each box indicate distribution median (black) and n value (blue). Mean number of measurements per site was 220 for *BET*+/– and 163 for *BEV* sites, with 34 yr and 30 yr of data, respectively. Overall, decadal median values are +0.5 cm/yr for *BET+* and –0.6 cm/yr for *BET*–, considerably higher than most documented long-term, basin-averaged rates of erosion in mountain belts and short-term rates derived from suspended sediment yields (Kirchner et al., 2001) (ASL—above sea level).

[1]GSA Data Repository item 2013163, supplementary methods, Figure DR1, and Tables DR1–DR4, is available online at www.geosociety.org/pubs/ft2013.htm, or on request from editing@geosociety.org or Documents Secretary, GSA, P.O. Box 9140, Boulder, CO 80301, USA.

the magnitude (not direction) of decadal trends in riverbed elevation (both *BET*+ and *BET*–) is correlated with *DA*, Q_{50}, and Q_{90} (Fig. 1D; Fig. DR1), indicating a positive first-order relationship between the magnitude of *BET* and *BEV* and the volume of erodible alluvial fill (and thus scour depth) in progressively larger basins, regardless of network structure, basin physiography, runoff-generating mechanisms, or geologic materials.

Imprint of Climate

Building on prior work (Molnar et al., 2006; Pitlick, 1994; Tucker and Slingerland, 1997), we investigate the hypothesis that the flood events capable of inducing long-term change in riverbed elevation are largely a function of regionally determined *P* variability (Tucker and Bras, 2000). Climatically, this should be expressed in *Q* distributions as a ratio between high and mean flows. Thus, we computed Q_{90}/Q_{50}, or high flow variability, as a proxy for riverbed shear-stress distribution above a sediment transport threshold (Baker, 1977; Molnar et al., 2006; Pickup and Warner, 1976). We truncated this distribution at Q_{90} in order to minimize the influence of rare floods, and normalized *BET*+ and *BET*– values by the median depth from all *Q* measurements for each site, in order to investigate the climatic expression of *Q* on *BET* in a scale-independent way.

We find monotonically increasing relationships between depth-normalized *BET*+ and *BET*– and Q_{90}/Q_{50} (Fig. 2A), consistent with the idea that high flow variability controls sediment flux and morphologic change (Molnar et al., 2006; Tucker and Slingerland, 1997). Bed material flux occurs at ~Q_{50} (depending on climatic regime) as transport of the active sediment layer (Emmett and Wolman, 2001), and on the high end of the *Q* range as reshaping of the channel (Pickup and Warner, 1976). Therefore, we may assume that stations with higher Q_{90}/Q_{50} are prone to more frequent channel scour/fill, subject to constraints of an erodible riverbed, leading to greater potential for *BET*. Stations with high Q_{90}/Q_{50} and with positive Q_{90}/Q_{50} trends might be expected to have greater potential for *BE* trends. [There are no systematic bed-material data available to test the importance of grain size on *BET* and *BEV*, but proxy data on basin lithology correlate poorly with *BEC* (Fig. DR1), suggesting that rock type is not a first-order control on *BE* in alluvial channels.]

Although the relationship between rates of *BET*+/– and mean*P* is poorly defined (Fig. DR1), we find that mean*P* has an indirect influence on *BEC*. Q_{90}/Q_{50}, which is directly correlated with *BET*+ and *BET*– (Fig. 2A), is inversely proportional to mean*P* (i.e., Q_{90}/Q_{50} decreases systematically with increasing mean*P*) (Figs. 2B and 2C), highlighting the variable expression of climate in *Q* distributions. Dry regions have higher Q_{90} for the same Q_{50} than wet regions, consistent with prior work (Turcotte and Greene, 1993). However, even though Q_{90}/Q_{50} is typically lower at wetter sites (mean*P* > 1000 mm), Q_{50} is apparently sufficiently high above the threshold for sediment flux at these sites to enable high values of change.

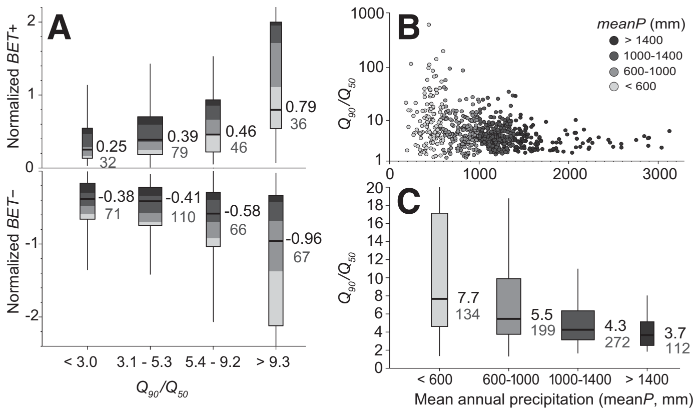

Figure 2. A: Bed elevation trend (*BET*) normalized by the median flow depth from all streamflow measurements at each site versus high-flow variability ratio (Q_{90}/Q_{50}; annual *Q* percentiles, where *Q* = discharge). B and C: Mean annual precipitation (mean*P*) versus Q_{90}/Q_{50}. Symbols as in Figure 1D.

Impacts of Changes in Climate

Given the monotonic relationships between Q_{90}/Q_{50} and *BET*/*BEV*, and the corresponding inverse one between Q_{90}/Q_{50} and mean*P* (Fig. 2), we question the potential impact of observed changes in *P* and *Q* on *BET* and *BEV*. To remove any scale bias we computed percentage changes in *Q*, and we find significant *Q*trends at 26% of sites for which sufficient data exist (n = 187) that reflect documented *P*trends for the contributing watersheds (Mitchell and Jones, 2005). *Q*trends (Q_{50} and Q_{90}) increase with *P*trend to 2 mm/yr, and negative *P*trends are associated with concomitant declining trends in both *Q* percentiles (Fig. 3A). We also find that moderately positive *P*trends (1–2 mm/yr) are disproportionately expressed as increasing *Q*trends in relatively dry watersheds, or those with high values Q_{90}/Q_{50}, yet the highest *P*trends are not associated with the highest *Q*trends (Fig. 3A), suggesting there is asymptotic buffering of extreme changes in rainfall in wet basins (e.g., via floodplain storage).

Furthermore, trends in both *Q* percentiles (Q_{50}trend and Q_{90}trend) are inversely related to mean*P*, such that *Q*trend strength progressively

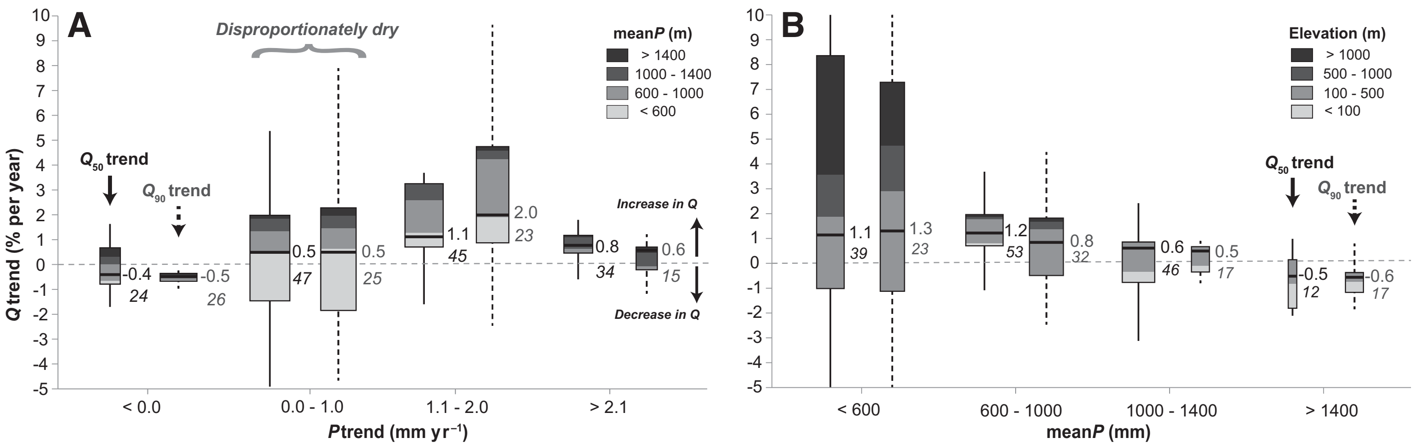

Figure 3. Annual percentage change in temporal trend in annual *Q* (i.e., discharge) percentiles (Q_{50}trend and Q_{90}trend) normalized by median value of *Q* for each site. A: Versus annualized trend in mean precipitation (*P*trend). B: Versus mean annual precipitation (mean*P*). Within each *x*-axis category, Q_{50}trend is represented by left box and Q_{90}trend by right box. Symbols as in Figure 1D. We find a similar number of sites with increasing and decreasing Q_{90}trends (n = 47:42), but twice as many sites with increasing Q_{50}trends (n = 103:49).

declines in increasingly wet watersheds (Fig. 3B). This translates as an overall increase in Q in dry regions (median +1.1% and +1.3% per year for Q_{50}trend and Q_{90}trend, respectively, for sites with mean$P$ < 600 mm), and a modest decrease at the wettest locations (median –0.5% and –0.6%, respectively, for those with mean$P$ > 1400 mm), compared to the overall median value of either Qtrend across all climates (~+0.7%). However, decreases in Q_{50} are also disproportionately expressed at dry sites and at high elevation (>1000 masl), and decreases in Q_{90} prevail disproportionately at wet sites (>1400 mm) (Fig. 3B). These disproportionate results suggest the complexity of climate change expression in streamflow.

How might such Qtrends affect bed elevation? Because the data sets are concurrent, it is challenging to discern cause from effect, especially if bed elevation response lags the climate change. However, we find that trends in both Q and *BE* occur at 18% of sites with Q data, and that marked increases and decreases (>+1% and <–1%) in Q_{50} and Q_{90} are both associated with higher rates of *BEC* (Table DR3 in the Data Repository). Humidification in both dry and wet climates is expressed through increasing trends in Q_{50} and/or Q_{90} (increases >+1%; Q_{50}trends are more common). The upward shift in median Q, which is typically greater than increases in Q_{90}, is consistent with decreasing Q_{90}/Q_{50}. This increases the flow duration above a fixed sediment flux threshold, potentially rendering the flow regime more geomorphically effective, regardless of small contractions in Q_{90}/Q_{50}. Aridification is typically expressed through declines in Q_{90}, yet expansion of Q_{90}/Q_{50} in dry and wet regions, and corresponds to high rates of *BEC* (see the Data Repository). At wet sites, one would expect aridification to reduce the already high number of sediment flux days, yet a small change is unlikely to have a major impact on *BET* and *BEV*, unless a transport/storage threshold is crossed. At arid sites it is plausible that changes in the flood frequency-magnitude relationship, especially in Q_{90}, may affect *BEC* over much longer time scales, since bed elevation is essentially a function of larger and less frequent channel-shaping floods, but further work is required to clarify this effect.

CONCLUSIONS

We document climatically influenced decadal *BE* trends and variability across the continental United States. We demonstrate that Q variability varies by climatic regime proportional to *BEC*. Climate change, affecting the variability in high Q in contrasting climates, may act to redistribute the prevailing signature of climate in alluvial riverbeds. We find nonstationarity in *BE* at most sites, which has important implications for assessing flood risk, aquatic habitat, and stability of infrastructure, as well as for channel sediment budgets and longitudinal profiles, and the overall impact of climate change on fluvial systems.

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