Extensive Frost Weathering Across Unglaciated North America During the Last Glacial Maximum

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Abstract In unglaciated terrain, the imprint of past glacial periods is difficult to discern. The topographic signature of periglacial processes, such as solifluction lobes, may be erased or hidden by time and vegetation, and thus their import diminished. Belowground, periglacial weathering, particularly frost cracking, may have imparted a profound influence on weathering and erosion rates during past climate regimes. By combining a mechanical frost-weathering model with the full suite of Last Glacial Maximum climate simulations, we elucidate the meters-deep magnitude and continent-spanning expanse of frost weathering across unglaciated North America at ~21 ka. The surprising extent of modeled frost weathering suggests, by proxy, the broad legacy of diverse periglacial processes. Complementing previous studies that championed the role of precipitation-driven changes in Critical Zone evolution, our results imply an additional strong temperature control on surficial process efficacy across much of modern North America, both during glacial periods and modern climes.

Plain Language Summary Hillslopes and rivers are shaped by both modern and past climates. Climate and ecosystems drive how fast bedrock weatheres, soil is produced, sediment moves downslope and rivers respond. We know that during glacial periods, unglaciated regions beyond the reach of ice sheets were much colder than today. By applying a frost-weathering model (which predicts changes in porosity, and thus rock damage, due to growth of ice lenses) to Last Glacial Maximum paleoclimatic simulations, we demonstrate the surprising extent and magnitude of periglacial processes ~21,000 years ago across much of North America. No matter which paleoclimate simulation one chooses, the results point to just how extensive frost weathering processes were during the last glacial period, and by extension, multiple glacial periods of earth’s history. Based on the widespread occurrence of glacial-period frost weathering over meter-scale depths, we suggest that past cold climates have had a significant impact on modern landscapes, both through lingering impact on subsurface pathways for water and thus chemical weathering, and the rock damage that contributes to the rate at which rock disaggregates into sediment and potential instability due to nonsteady rates of hillslope and river processes.

1. Introduction

Copious paleoenvironmental records in mid-latitude unglaciated regions establish profound differences between modern temperate and past colder climate conditions (e.g., Bartlein et al., 2014; Lowe & Walker, 2015 and citations within). Recent advances in coupling cosmogenic-derived erosion rates and frost-weathering models from both modern (Delunel et al., 2010; Mair et al., 2020) and paleo-settings (Marshall et al., 2017) supported by paleoclimate simulations (Marshall et al., 2015) have illuminated the potential for pervasive periglacial processes operating across unglaciated landscapes during glacial intervals. In these cold paleoclimate settings, isotope-derived erosion rates are commonly 2–4x faster during glacial intervals than during interglacial intervals (Dosseto & Schaller, 2016; Schaller et al., 2002; Scherler et al., 2015). These observations suggest a substantial role for past climates in constructing modern subsurface hillslope architecture (Anderson et al., 2013; West et al., 2019) and thus influencing modern Critical Zone processes. The Critical Zone extends from where vegetation intersects the atmosphere to the lowermost extent of weathered bedrock. As fresh bedrock is exhumed toward the surface and intersects with water and organisms, rock properties such as porosity, permeability and fracture density control Critical Zone attributes including: the available surface area for geochemical reactions (Li et al., 2017), pathways for hydrologic routing (Hahm et al., 2019), fracture space for roots, and the mobile fraction of weathered material (Sklar et al., 2017).
These attributes are determined not only by present-day processes but are predicated on the “ghosts” of past processes embedded in the weathered rock below the surface. Growing evidence of past periglacial processes active in mid-latitude terrain suggests a reconsideration of the style, magnitude, and extent of change in landscape processes during glacial intervals and consideration of their potential import for influencing modern processes.

In modern temperate settings, based on field evidence such as frost-shattered rock, solifluction lobes, blockfields, polygonal cracking, and sand wedges, previous workers have mapped discrete or regional areas of cold climate artifacts beyond the maximum extent of northern hemisphere ice sheets (Lindgren et al., 2016; Vandenberghe et al., 2014). This archive of past periglacial locations is expanding rapidly with the ready availability of remote satellite imagery (i.e., Google Earth) and lidar (Del Vecchio et al., 2018; Gao, 2014). The distribution of polygonal cracks and sand wedges, considered the most diagnostic of permafrost features, has been used to delineate the extent of permafrost in North America (French, 2007; Vandenberghe et al., 2014) as well as to estimate past climate conditions (Mears, 1981). These permafrost maps can be biased by the exclusion of heavily forested and fast eroding areas, where surface expressions of past periglacial processes are either hidden or have been eroded away. Additionally, because polygonal cracking arises from extensive periods under deeply frozen conditions (i.e., with temperatures ranging from ∼−5 °C to −15°C; Vandenberghe et al., 2014), regions undergoing warmer periglacial processes such as frost-driven bedrock weathering (i.e., with temperatures extending up to −3°C or warmer; Rempel et al., 2016; Walder & Hallet, 1985) or solifluction, which can occur when temperatures hover within a few degrees of the melting point (Matsuoka, 2001), are likely to be excluded from these mapped permafrost zones (Mears, 1981; Vandenberghe et al., 2014).

As cautioned by Bryan (1950), a bias toward the familiar and known (e.g., precipitation-driven landslides delivering sediment to rivers and influencing river form) can color our conceptual understanding and the associated hypotheses we formulate for discerning past processes. Despite abundant paleoenvironmental evidence of past cold conditions (Bartlein et al., 2011; Jackson et al., 2000), the influence of cold temperatures on surface and near-surface processes in mid-latitude unglaciated regions has at times been discounted in favor of interpretations focused on changes in precipitation (Bull, 1991; Perron, 2017; Personius, 1993). However, while water supply is an important factor in many weathering processes, in the periglacial environments that characterized extensive tracts of North America in the past, key processes such as frost weathering are much more sensitive to temperature and temperature variations.

Here, we combine a physically based, state-of-the-art frost-weathering model with downscaled paleoclimate simulations to explore the potential extent of associated cold climate processes across the unglaciated extent of North America during the Last Glacial Maximum (LGM) ∼21 ka. As we are exploring the potential for frost weathering at the continental scale, we ignore regional differences in lithology, snow cover, aspect and other factors that can locally amplify or dampen frost weathering conditions, and instead focus this contribution on commonalities in our suite of model outputs.

2. Ice as a Weathering Agent Via Frost Cracking

Ice is an extremely effective weathering agent, capable of weakening and shattering its host rock (Anderson et al., 2013; Draebing & Krautblatter, 2019; Hales & Roering, 2007; Hallet et al., 1991). A fixed mass of water expands upon freezing and can thereby exert appreciable stresses against confining walls. However, while volumetric expansion is often invoked in discussions of frost weathering (Hallet, 2006), liquid mobility and the common occurrence of unsaturated conditions in porous rock severely limit the effectiveness of this mechanism (Hallet, 2006). Moreover, observations reveal that the most significant frost damage often occurs only once the temperature has dropped several degrees below that at which the latent heat signature is detected; this indicates that most liquid has already frozen, with any associated volumetric expansion taking place long before significant damage occurs (e.g., Draebing & Krautblatter, 2019; Jia et al., 2017). Instead, a spate of theoretical, experimental, and field results (e.g. Murton et al., 2006; Rempel et al., 2016; Walder & Hallet, 1985) give overwhelming support for ice weathering mechanisms that rely upon predicted and observed localized increases in ice mass beyond the capacity of the original undeformed, liquid-saturated pore space (sometimes referred to as segregated ice growth). These mass increases result from liquid pressure
Gradients that drive flow to freezing centers in near-surface pores (in both fresh and weathered bedrock). The associated increase in porosity can be accommodated by inelastic cracking and is thereby tied directly to physical rock weathering (Hallet et al., 1991; Murton et al., 2006).

Frost-driven cracking is sensitive to pressure exerted by ice against crack walls (which increases at colder temperatures), the permeability to liquid flow (which decreases at colder temperatures), and the magnitude of the temperature gradient (which induces the pressure gradients that cause liquid flow). As a result, the accumulation of frost damage depends on both the temperature and, importantly, the amplitude of seasonal temperature changes, which exerts a dominant control on the temperature gradient at decimeter to meter scale depths. Thus, factors that control the amplitude of the annual cycle of temperature such as proximity to ice sheets or large water bodies, insolation, and topography (e.g., mountains vs. valleys) can produce significant differences in the amount of bedrock damage between sites with similar mean annual temperatures (Marshall et al., 2015). In some locations the annual temperature amplitude is greater at lower elevations than at nearby higher elevations sites, which can result in elevated frost-weathering potential on some valley floors relative to the nearby colder mountains (Marshall et al., 2015). Also, because climate pattern seasonality may differ during glacial versus interglacial periods, it is vital to recognize that conceptual models of temperature distributions based on modern climate patterns do not always apply to glacial periods (Izumi et al., 2013).

Here, we model frost weathering by tracking changes in porosity that result from the liquid flow that enables ice growth to extend cracks. The modeled increases in porosity imply depth-integrated expansion on the order of $10^{-1}$ mm yr$^{-1}$ concentrated in the uppermost decimeter to meter (Rempel et al., 2016), which yield typical strain rates of order $10^{-4}$ yr$^{-1}$ that are likely sufficient to significantly weaken rock over time (Eppes & Keanini, 2017; Hayes et al., 2019; Tuğrul, 2004). As porosity increases, surface area and permeability increase, allowing for greater weathering due to chemical reactions, and increased diffusivity due to enhanced pore connectivity (Navarre-Sitchler et al., 2011). Thus, as frost weathering advances fractures, the rock becomes increasingly susceptible to both physical and chemical weathering (Holbrook et al., 2019) and associated disaggregation. This suggests that as the magnitude of ice-driven porosity increases, so too does rock damage.

### 3. Methods

#### 3.1. Experimental Design

To evaluate the distribution and magnitude of depth-integrated annual frost-weathering damage ~21 ka across our North America study region, we used output from a suite of available LGM paleoclimate simulations to drive an existing physically based frost-weathering model that tracks the ice mass increases that cause bedrock porosity ($n$) to change with depth ($z$) as frost cracking occurs (Rempel et al., 2016). After downscaling the climate simulations to a 10-km grid (described below) and masking for LGM alpine glaciers and ice sheet extent, we used the mean annual temperature and the amplitude of the annual cycle for each grid cell to calculate (a) changes in porosity ($n$) over decimeter intervals down to 10 m ($z$), (b) the depth-integrated change in porosity ($\lambda$), and (c) the median swell depth ($d$). Combined, these simulations provide a regional to continental-scale view of the extent and degree of 1 year of frost-driven changes in porosity across unglaciated North America during the LGM.

#### 3.2. The Frost-Weathering Model

In our model, as the temperature, $T$, drops, there is a corresponding increase in the undercooling $\Delta T$ defined as the departure from the normal bulk melting temperature $T_m$. The undercooling is directly proportional to the difference between the ice pressure exerted against crack walls and the reduced pressure in residual thin films that facilitates ongoing liquid supply and progressive increases in ice mass (Dash et al., 2006). Cracking occurs when rock is colder than the upper temperature limit to frost weathering $T_c$, such that once $\Delta T > \Delta T_c$, the ice pressure exerted against the rock matrix is sufficient to propagate cracks. The water supply is limited by the permeability $k$ (m$^2$), which decreases with $\Delta T$ as ice clogs the pore space; this variation is modeled by a power law.
\[ k = k_c \left( \frac{\Delta T}{\Delta T_c} \right)^\alpha \]  

(1)

where \( k_c \) is the permeability at \( \Delta T_c \) and \( \alpha \) is an empirical exponent set to 4, a typical value based on power law approximations for permeability in subzero conditions validated in fine-grained soils (Andersland & Ladanyi, 2004). Since \( k \) is affected so dramatically by \( T \) changes, we ignore comparatively minor changes in viscosity, which have a negligible influence on the model results. Unlike many previous frost cracking models (Anderson, 1998; Hales & Roering, 2007; Marshall et al., 2015), we do not include an explicit low-temperature limit to frost damage (typically described as the lower bound of the frost-cracking window). Rather, as temperatures decrease, the corresponding decrease in permeability effectively retards the supply of liquid water necessary for continued ice growth and thus it is the reduced permeability that sets the lower temperature boundary for significant ice-induced changes in porosity over time.

We use Darcy’s law to model water flow with the undercooling-dependent permeability from Equation 1, driven by liquid pressure gradients that are also controlled by the undercooling because of the thermodynamic considerations outlined above. With the ice pressure along expanding crack surfaces controlled by geometry and rock strength, mass balance implies that the annual change in porosity satisfies

\[ \Delta n(z) = \frac{D}{\Delta T_c} \left\{ \Delta T > \Delta T_{T_c,1\text{year}} \right\} \left( \frac{\Delta T}{\Delta T_c} \right)^{\alpha+1} \left( \frac{\partial T}{\partial z} \right)^2 \right\} dt \]

(2)

where, the diffusivity parameter \( D \) (m²/s) combines the restrictions on permeable flow, the threshold undercooling needed for crack propagation, and thermodynamic parameters, according to

\[ D = \frac{\alpha \rho L \Delta T}{T_{\text{m}} \mu} \]

(3)

where \( \rho \) is the ice density (kg/m³), \( L \) is the latent heat (kJ/kg), and \( \mu \) is water viscosity (mPa s). A useful metric for assessing lateral variations in the degree of frost damage is the depth-integrated annual increase in porosity (\( \lambda \)), defined by

\[ \lambda = \int \Delta n dz. \]

(4)

Ice-liquid phase equilibrium conditions, as described by the Clapeyron equation, ensure liquid water availability as long as ice is present. See Table S1 for a complete list of parameter values and typical ranges. An expanded model derivation is provided by Rempel et al. (2016).

### 3.3. Paleoclimate Simulations

We combined the physically based frost-weathering model described above with downscaled (10-km) LGM temperature simulations for our study area (Shafer et al., 2021). To best represent the range of potential LGM climates and thus frost weathering, as well as to highlight the communality in the distributions of simulated frost weathering, we used temperature simulations produced under the Coupled Model Intercomparison Project phase 5 (CMIP5)/Paleoclimate Modeling Intercomparison Project phase 3 (PMIP3) \( \text{lgm} \) (LGM) and \( \text{piControl} \) (preindustrial) experiments (Harrison et al., 2014) (terms in italics are the database field names) from the following climate models: CCSM4, CNRM-CM5, COSMOS-ASO, FGOALS-g2, GISS-E2-R, IPSL-CM5A-LR, MIROC-ESM, MPI-ESM-P (OA and OAC), and MRI-CGCM3 (Table S2), and combined these simulations to generate the ensemble average. To downscale the mean monthly temperature data, we re-gridded 1961–1990 30-year monthly mean temperature data from the CRU CL 2.0 10-min data set (New et al., 2002) to the 10-km equal-area grid using topographic local lapse-rate adjusted interpolation (Praskievicz & Bartlein, 2014). The downscaled CRU CL 2.0 data were used as our “modern” data set. We then calculated \( \text{lgm} \) minus \( \text{piControl} \) long-term mean differences (or “anomalies”), applied the long-term mean differences for individual models to the present-day CRU CL 2.0 data on the 10-km grid, and also calculated the ensemble, or multimodel, averages. We fit a sine curve to the downscaled monthly mean surface temperatures from the paleoclimate simulations, enabling us to extract spatial variations in mean
annual temperature MAT and annual temperature amplitude A. We used this simplified two-parameter fit as the boundary condition for our subsurface temperature calculations, performed at 1 s resolution over an annual cycle. While the single harmonic fit serves our primary goal of highlighting the dominant climatic controls on periglacial processes at the continental scale, we note that more detailed analyses that focus on the effects of local annual temperature variations may often be represented significantly better using two or more harmonics.

### 3.4. Swell Depth

The shape of the depth profile for annual frost-driven porosity varies, depending on the combined influence of MAT and A. For example, zones of high-intensity frost damage can occur primarily in the upper meter, sharply spike at 2 m, remain essentially uniform down to 4 m, or exhibit a nearly linear decline with depth extending beyond 6 m from the surface (Anderson et al., 2013; Hales & Roering, 2007; Rempel et al., 2016). We follow Rempel et al. (2016) in using the term “swelling depth,” defined as the depth from the surface over which half the total expansion due to increases in porosity has occurred. In other words, it is the median depth for the integrated annual increase in porosity such that:

$$\int_0^d \Delta(z) \, dz = \frac{\lambda}{2}$$

Thus, the swell-depth metric, d, represents the zone of intense rock damage in a manner that is more meaningful than alternatives such as the predicted maximum depth of frost weathering, which does not reflect dramatic declines in the degree of frost damage with depth.

### 3.5. Model Considerations

We simplify our treatment of ground temperatures by neglecting the effects of latent heat when approximating the temperature profiles. While damage is reduced under unsaturated conditions, whereas damage is enhanced by the high temperature gradients that accompany increased radiative transfer (Rempel & Rempel, 2019), we make no attempt to take such factors into account. Our focus on gauging regional variations in damage over annual cycles motivates our simplified approach in approximating surface temperatures with the output of the paleoclimate simulations described below, and assuming the conditions necessary for full saturation throughout. Our model treatment addresses the important end-member case of frost cracking in exposed, saturated bedrock. More generally, complicating factors, such as hillslope aspect, cloud cover, tree shading, snow and soil insulation, can influence the temperature at the bedrock surface and modulate diurnal temperature swings in the uppermost millimeters to decimeters of bare and soil-covered rock. Moreover, by applying a single set of physically based parameters (Table S1) from the mid-range of measured values, we ignore lithologic-driven variabilities in rock strength and permeability, which are important when considering rock-driven cracking for individual rock masses. For the purposes of this continental-scale modeling investigation, we do not include these complexities and focus instead on regional patterns that are expected to be most sensitive to the degree of frost penetration over the decimeter-to-meter-scale depths that are responsive to annual temperature swings. The strong relationship between local temperature gradients and the accumulation of frost-driven weathering that emerges from our analysis suggests that bedrock up to a meter or two below the ground surface should be subject to significant frost cracking.

### 4. Results

#### 4.1. Significance of Mean Annual Temperature and Annual Temperature Amplitude

The generally greater amplitude of the annual cycle of temperature at the LGM (relative to present) was a consequence of the greater high-latitude/low-latitude and land-ocean contrasts in cold climates relative to warm ones (Izumi et al., 2015). Our model of depth-integrated changes in porosity (λ) over one year illuminates the combined effect of mean annual temperatures (MAT) and annual temperature amplitude (A) in controlling the magnitude of frost-driven increases in porosity (Figure 1). Even modest decreases in MAT combined with increases in A will result in significant increases in frost weathering. The liquid pressure
gradients that supply ice growth can be related to the temperature gradients at the upper and lower boundaries of the active frost weathering region, while thermal diffusion causes the amplitude of the sinusoidally varying temperature gradient to decay with depth. It is this control of heat flow on liquid transport that gives rise to the bowed shapes of the isolines in Figure 1 (see Rempel et al., 2016).

4.2. Predicted Magnitude and Extent of Frost Weathering

To contextualize frost-weathering extent during the LGM, we compared the modeled modern frost-cracking extent with the modeled LGM extent generated from the individual LGM climate simulations and the ensemble. We focus on comparing the spatial distribution of frost-weathering for the continental region south of the North American LGM ice sheets. Our results suggest significant frost weathering occurred across broad swaths of unglaciated North America for the 10 individual simulations, and for the ensemble (Figure 2, Figure S1). Surprisingly, several simulations suggest frost weathering as far south as 30°N in the interior of the North American continent. The combination of proximity to the ice sheets and distance from the buffering influence of the oceans causes LGM frost weathering to extend farther southward in the interior regions, similar to the extensive southward extent of the interior portions of the ice sheet. This zone of high frost weathering at the center of the continent is driven by large A and/or MAT within a few degrees of 5 °C (based on our continental-scale model parameter choices, Figure 1), where conditions facilitate liquid transport (Rempel et al., 2016). Individual paleoclimate simulations generate annual increases in λ as high as 0.84 mm (Figure S1, Marshall et al., 2020). While the frost weathering magnitude varies among simulations, the continental-scale patterns remain consistent.

Our simulations suggest that during the LGM, frost-weathering conditions would have extended over an area >2x larger than present (∼5,320,000 km² vs. ∼2,230,000 km²). Notably, the predicted cold-climate-driven frost-weathering extent is significantly greater than the LGM permafrost extent (Figure 2b) (Lindgren et al., 2016, and citations within) and extends to latitude 25°N (Texas) with significant frost-weathering as far south as 30°N (Oklahoma and Arkansas). Excluding Beringia, our predicted area of North American frost weathering is ∼3.5 times greater than the area estimates for permafrost extent (∼5,320,000 km² vs. ∼1,580,000 km²).

Our models also suggest that frost weathering extended to geomorphically significant depths below the surface (Figure 2c and Figure S2). Given that κ is a function of the material’s thermal conductivity, specific heat capacity and density, with typical values of κ ranging from 0.1 to 2 (mm²/s), there is potential for the swell depth to reach as much as 5 m below ground, which would imply significant frost-driven weathering at depths that are geomorphically significant over time scales on the order of 10^3–10^5 years depending on the erosion rate.

5. Discussion

Hillslopes occupy over 90% of unglaciated land (Kirkby, 1978), and our results suggest a re-consideration of the importance of subfreezing temperatures in shaping landscapes in mid-latitude terrain due to the potential for changes in rock damage, frost-weathering-derived sediment production and solifluction-driven transport (Bovy et al., 2016; Marshall et al., 2017; Matsuoka, 1998, 2001; Murton & Belshaw, 2011). Though sparse, mapped periglacial artifacts such as solifluction deposits and rock streams are coincident with our predicted southern extent of frost weathering (French & Millar, 2014). Given the continental-scale decrease in temperatures (∼10°C–15°C colder), with colder summers and much colder winters than present (Bartlein et al., 1998; Tierney et al., 2020), the combined effect of decreased MAT and increased A during the LGM
Figure 2. Extent and magnitude of modern versus LGM frost weathering (a) Modern extent and magnitude of $\lambda$, the depth-integrated annual increase in porosity. Cities highlighted on Figure 1 delineated with letters. Cities: A—Atlanta, Georgia; D—Dallas, Texas; SL—St. Louis, Missouri; DC—Washington D.C. (b) Predicted $\lambda$ for the CMIP5/PMIP3 ensemble-average climate. LGM permafrost extent (Lindgren et al., 2016) included to showcase latitudinal differences between LGM permafrost versus predicted extent of frost weathering. (c) Predicted annual LGM swell depth ($d$) for the CMIP5/PMIP3 ensemble. Frost cracking model output for all available LGM paleoclimate simulations can be found in supporting information Figures S1 and S2. Ice sheet extent from Dyke et al. (2003).
(and by extension during other glacial intervals) would have led to substantial increases in frost-weathering intensity and extent. It is worth noting that while the ensemble simulations predict a maximum increase in \( \lambda \) of 0.57 mm, paleoclimate simulations tend to generate warmer temperatures than the paleo-evidence (Harrison et al., 2014).

The legacy of cold-climate processes may have long-term implications for Critical Zone and landscape evolution. The extent of this relict impact depends on its extent and magnitude as well as the pace of erosion which controls preservation potential. Thus, damage from past climates may persist for millennia in slow eroding settings. In fast-eroding settings, one can observe hillslopes that plunge to V-shaped valleys carved by fluvial and debris flow processes. The signature of climate change in these landscapes is often difficult to discern even though the rock below the surface may still record frost-driven weathering at depths exceeding those of modern physical weathering mechanisms (e.g., tree-root wedging: Roering et al., 2010) or cyclic thermal cracking (Eppes et al., 2016). Frost-driven increases in porosity, superimposed on tectonic and/or topographic rock damage, will likely accelerate weathering and provide pathways for water and associated chemical weathering as well as egress for deep roots. Thus, in locations where the bedrock has been damaged due to past influence of frost, we would predict accelerated weathering and susceptibility to erosion relative to regions with similar modern climates and less periglacial preconditioning (Anderson et al., 2013; West et al., 2019).

Importantly, regions with large swell depths may be distinct from regions with the most intense frost-driven increases in porosity (Figure 2, Figures S1 and S2). Thus, LGM imprint may persist in both slow eroding regions where frost weathering was intense and fast eroding regions where the swell depth was deep. While the paleoclimate simulations imply intense frost weathering centered in the midcontinent, the spatial extent of swell depths approaching 2 m encompasses a latitudinal swath across the unglaciated North American continent, mirroring the shape of the ice sheets, such that all models predict a wide, equatorial-bowed band of deep weathering (Figure 2, and Figure S2). This predicted swath of deep weathering extends as far south as 38°N across central North America. The deepest swell depth zone does not coincide with the mapped permafrost extent, as temperatures near the ice sheet were too cold for frost weathering during the winter months (Bartlein et al., 1998; Tierney et al., 2020) reflecting the limiting effects of thermal diffusion. More generally, deeper and more intense frost weathering occurs in permafrost regions compared to areas with seasonally frozen ground (Murton et al., 2006), suggesting either a lack of preserved diagnostic permafrost features and/or permafrost conditions in these regions beyond the mapped periglacial extent.

Hillslope weathering influences not only hillslope function (e.g., soil thickness, subsurface void space for water storage, and the reactive surface area for chemical weathering; Caves Rugenstein et al., 2019), but the grain size distribution supplied to channels, a first order control on river incision and aquatic ecosystem function and distribution (Sklar et al., 2017). Large block delivery to channels, observed in formerly periglacial settings such as Pennsylvania (Del Vecchio et al., 2018), and perhaps as far south as northern Arkansas (Figure 2), can limit river incision, serving as a rate-limiting process in modern slow eroding settings (Shobe et al., 2016; Thaler & Covington, 2016) where our results suggest significant amounts of frost weathering during the LGM.

While this study is aimed at characterizing the extent of LGM frost weathering, low-temperature conditions necessary for frost weathering and other periglacial processes develop during the onset of cold climate conditions, thousands of years before glacial maxima (Lisiecki & Raymo, 2005). As temperatures decline during transitions from warm interglacials into glacial periods, conditions sufficient to generate frost weathering likely developed across a significant swath of North America. Paleoenvironmental and erosion rate data from lake cores extracted from Little Lake, Oregon, a site 400 km to the south of the maximum extent of the Cordilleran ice sheet, suggest subalpine conditions (similar to southeastern Alaska) and vigorous periglacial processes developed by \( \sim 35 \) ka, thousands of years before Marine Oxygen Isotope Stage 2 (the LGM), which began at \( \sim 29 \) ka (Grigg et al., 2001; Lisiecki & Raymo, 2005; Marshall et al., 2017). By extension, this result implies that over 1.6 million km\(^2\) of North American landmass (assuming a continental width of \( \sim 4,000 \) km and a 400 km swath south of the ice sheet) could have been subject to frost weathering for \( > 10 \) ky (from \( \sim 33 \) to 18 ka).
Although much of western North America was wetter than present during glacial intervals (Marshall et al., 2015), this does not preclude frost weathering and other periglacial processes from occurring during glacial period winter seasons in these regions. Thus, conceptual models such as those attributing river terrace formation throughout western North America to increased precipitation during glacial periods may be missing temperature as a key contributor to increased sediment production, transport and associated river incision or aggradation (Bull, 1991). Given the breadth of continental area with conditions suitable for frost weathering and other rate-changing periglacial processes (Figure 2) (French, 2007; Matsuoka, 2001), our results suggest the potential for highly transient behavior in these landscapes. Additionally, variations in the vigor and depth of LGM frost weathering may precondition a modern hillslope to weather faster than might be expected if only considering modern climatic controls. Thus, millennial-scale fluctuations may be the norm rather than the exception, as hillslope response to perturbations is on the order of ~50–100 ky (Roering et al., 2001), leaving scant time for steady-state conditions over the Pleistocene, with climate fluctuations between warm and cold periods occurring on timescales of ~40 ky –100 ky (Lisiecki & Raymo, 2005).

Although changes in porosity due to frost weathering do not necessarily imply increased erosion, efficient rock damage is a necessary precursor to increased soil production and transport reflected in paleo-erosion records (Delunel et al., 2010; Dosseto et al., 2012; Marshall et al., 2017; Schaller et al., 2002). As such, our findings may have broad application for interpreting the influence of periglacial processes past and present. In addition, local conditions, such as snow or sediment cover or terrain aspect, can enhance or dampen frost weathering (Anderson et al., 2013; Girard et al., 2013). Consideration of these factors does not diminish our findings but rather helps to clarify local-scale heterogeneity within broader patterns of our continental-scale simulations.

6. Conclusions

Our results, based on paleoclimate model simulations, indicate that frost processes in unglaciated mid-latitude terrain were likely pervasive across broad swaths of unglaciated terrain. The results suggest that periglacial processes may have profoundly modified mid-latitude unglaciated terrain, even in areas where frost weathering was moderate and/or precipitation played a major role during glacial periods. In detail, frost damage is expected to be sensitive to variations in bedrock type, soil depth and other factors, that we have not considered here. Local studies, tied to regional paleoclimatic data and site-specific parameters, will enable more advanced hindcasting of weathering and erosion rates (e.g., Marshall et al., 2015). By integrating a subsurface view of past weathering potential into our conceptual and numerical models, we can better predict how the weathering ghosts of climates past are exhumed from eroding and uplifting hillslopes and transported from hill to valley.

Data Availability Statement

Downscaled CMIP5/PMIP3 climate simulations output are available at https://doi.org/10.5066/P9KC0L47. Modeled values for predicted frost weathering across North America 21 ka based on all available downscaled CMIP5/PMIP3 climate simulations can be found at https://doi.org/10.6084/m9.figshare.12786638.v1. Frost cracking model code is available from J. A. Marshall on request. Any use of trade, firm, or product names is for descriptive purposes only and does not imply endorsement by the U.S. Government.

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