Catastrophic rock avalanche 3600 years BP from El Capitan, Yosemite Valley, California†

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ABSTRACT: Large rock slope failures from near-vertical cliffs are an important geomorphic process driving the evolution of mountainous landscapes, particularly glacially steepened cliffs. The morphology and age of a 2·19 × 10^6 m^3 rock avalanche deposit beneath El Capitan in Yosemite Valley indicates a massive prehistoric failure of a large expanse of the southeast face. Geologic mapping of the deposit and the cliff face constrains the rock avalanche source to an area near the summit of ~8·5 × 10^4 m^2. The rock mass free fell ~650 m, reaching a maximum velocity of 100 m s⁻¹, impacted the talus slope and spread across the valley floor, extending 670 m from the base of the cliff. Cosmogenic beryllium-10 exposure ages from boulders in the deposit yield a mean age of 3·6 ± 0·2 ka. The ~13 kyr time lag between deglaciation and failure suggests that the rock avalanche did not occur as a direct result of glacial debuttressing. The ~3·6 ka age for the rock avalanche does coincide with estimated late Holocene rupture of the Owens Valley fault and/or White Mountain fault between 3·3 and 3·8 ka. The coincidence of ages, combined with the fact that the most recent (AD 1872) Owens Valley fault rupture triggered numerous large rock falls in Yosemite Valley, suggest that a large magnitude earthquake (≥M7.0) centered in the south-eastern Sierra Nevada may have triggered the rock avalanche. If correct, the extreme hazard posed by rock avalanches in Yosemite Valley remains present and depends on local earthquake recurrence intervals. Published in 2010 by John Wiley & Sons, Ltd.

KEYWORDS: landslide; rock fall; seismic trigger; cosmogenic nuclide dating

Introduction

Large rock slope failures are among the most effective of the processes shaping mountainous landscapes. Although these events have relatively low frequencies, they move large masses of material over long distances, performing in geologically instantaneous time the amount of work accomplished over millennia by more frequent but smaller magnitude surface processes. Large rock slope failures are responsible for large proportions of mountain erosion, and are thus important events in controlling long-term rates of landscape evolution in mountainous regions (Hovius and Stark, 2007). Because earthquakes and climate events (in the form of glacial retreat or intense precipitation events) are common triggers of large rock slope failures, their study can elucidate the tectonic and climatic forces controlling hillslope adjustment to topographic relief (Korup et al., 2007). Large rock slope failures from near vertical cliffs, under-represented in geologic study, are an important means by which glacially-modified landscapes evolve.

Yosemite Valley is a ~1 km deep canyon carved into granitic rocks of the Sierra Nevada batholith (Bateman, 1992). The valley has fluvial origins, but was substantially deepened and widened by Pleistocene glacial erosion (Matthes, 1930; Gutenberg et al., 1956; Huber, 1987). Because the granitic rocks in the Yosemite region are relatively massive with high rock mass strength values, they support extremely steep, even overhanging, rock faces such as El Capitan, Half Dome, Royal Arches, and Cathedral Rocks. Yosemite Valley was most recently glaciated during the Last Glacial Maximum (LGM), locally termed the Tioga Glaciation, which lasted from ~28–14 ka in the Sierra Nevada (Matthes, 1930; Huber, 1987; Phillips et al., 1996, 2009; Clark et al., 2003).

The glacially steepened cliffs of Yosemite Valley experience frequent rock falls. A database extending from AD 1857 to the present documents over 600 rock falls or other forms of slope movement, such as rock slides and debris flows, in Yosemite National Park (Wieczorek and Snyder, 2004, and subsequent observations). The vast majority of these events were rock falls and rock slides in Yosemite Valley, with individual rock fall
volumes up to ~600,000 m$^3$. Failures often occur as planar or slab failures along surface-parallel sheeting joints. Analysis of the rock-fall database reveals a range of triggers, including precipitation, freeze–thaw, and thermal stresses, although >50% of documented rock falls lack recognized triggering mechanisms (Wieczorek and Jäger, 1996; Wieczorek and Snyder, 2004).

Earthquakes have triggered at least 20 historic rock falls in Yosemite Valley, for a cumulative volume of ~67,400 m$^3$ (Wieczorek and Jäger, 1996; Wieczorek and Snyder, 2004). The 25–27 May 1980 earthquakes centered near Mammoth Lakes, with magnitudes as great as M 6.5, triggered thousands of rock falls and rock slides in the Sierra Nevada (Harp et al., 1984) and at least nine rock falls in Yosemite Valley (Wieczorek and Jäger, 1996; Wieczorek and Snyder, 2004). The 26 March 1872 Owens Valley earthquake, one of the largest historic earthquakes in California (M 7.5; Beanland and Clark, 1994), triggered at least five large (up to 36,000 m$^3$) rock falls in Yosemite Valley (Wieczorek and Snyder, 2004), which is located 180 km north-west of the rupture. The conservationist John Muir famously described one of the resulting rock falls:

> At half past two o’clock of a moonlit morning in March, I was awakened by a tremendous earthquake . . . The shocks were so violent and varied, and succeeding one another so closely . . . it seemed impossible that the high cliffs of the Valley should escape being shattered . . . Then suddenly . . . there came a tremendous roar. The Eagle Rock on the south wall, about half a mile up the Valley, gave way and I saw it falling in thousands of great boulders . . . pouring to the Valley floor in a free curve luminous from friction, making a terribly sublime spectacle – an arc of glowing, passionate fire, fifteen hundred feet span . . . (Muir, 1912).

Wieczorek (2002) estimated that this rock fall, originating from the southern slope of Yosemite Valley between Glacier Point and Sentinel Rock, had an approximate volume of 20,000 m$^3$.

Beyond historical timescales, longer-term rock-fall activity in Yosemite is reflected in the large volumes of talus that have accumulated beneath the cliffs since retreat of the LGM glaciers (Wieczorek and Jäger, 1996; Moore et al., 2009). Active talus slopes in Yosemite Valley are commonly 100 m thick (over 200 m thick in places) and extend horizontally 100 m or more beyond the base of cliffs. The distal edge of active talus slopes represents the limit that typical rock falls extend beyond cliff faces (Wieczorek et al., 1999).

At five locations in Yosemite Valley, rock debris has traveled far beyond the distal edge of active talus slopes and out onto the valley floor (Wieczorek et al., 1999). These deposits have been interpreted as rock avalanches based on their size and runout distances (Wieczorek et al., 1999; Wieczorek, 2002). Rock avalanches are massive slope failures consisting of mostly dry rock debris and generally characterized by distinctive morphologies, long runout distances relative to their fall heights, velocities of tens of meters per second, and volumes exceeding 1·0 × 10$^6$ m$^3$ (Shreve, 1968; Scheidegger, 1973; Hsu, 1975; Nicoletti and Sorriso-Valvo, 1991; Davies and McSaveney, 1999; Strom and Korup, 2006; Mitchell et al., 2007; Shea and van Wyk de Vries, 2008). Given their large size, rock avalanches are an important geomorphic process driving landscape evolution in Yosemite Valley. With nearly four million visitors to Yosemite National Park annually, rock avalanches also pose an extreme hazard that warrants additional investigation. Here we present research results on the magnitude, timing, and potential triggering mechanisms of a prominent rock avalanche beneath the near-vertical south-east face of El Capitan.

El Capitan Meadow Rock Avalanche

Rising 940 m above the floor of Yosemite Valley (mean elevation 1208 m), with a vertical to overhanging face up to 895 m tall, El Capitan is among the largest and most widely-recognized granitic rock faces in the world (Figure 1). Historically, there have been relatively few rock falls documented from El Capitan (Wieczorek and Snyder, 2004, and subsequent observations). However, this is at least in part due to the fact that rock-fall reporting rates are comparatively low in this undeveloped area. A large apron of active talus beneath El Capitan, ranging from 45 to 225 m high and extending horizontally outward from the cliff face 160–400 m, is evidence of extensive post-glacial rock-fall activity (Figures 2, 3).

Below the south-east face of El Capitan, a tongue-shaped deposit of rock debris extends well beyond the distal edge of the active talus slope (Figures 2 and 3). Matthes (1930) described this deposit as follows:

> . . . the most remarkable body of earthquake débris is that which lies in front of El Capitan – not the talus of blocks that slopes steeply from the cliff to the valley floor, but the much vaster hummocky mass, partly obscured by a growth of trees and brush, that sprawls nearly half a mile out into the valley. There can be no doubt that it is the product in the main of one colossal avalanche that came down from the whole height of the cliff face – probably the most spectacular rock avalanche that has fallen in the Yosemite Valley since the glacial epoch . . . the quantity of débris that fell in this stupendous earthquake avalanche is so great . . . that its removal doubtless altered appreciably the contour and appearance of El Capitan.

This deposit, termed the El Capitan Meadow rock avalanche (Wieczorek et al., 1999), is well resolved in a 1 m digital elevation model (DEM) and shaded relief model (Figure 2) created from filtered airborne LiDAR data. The rock avalanche deposit has a lobate morphology where it emerges from
Figure 2. Shaded relief model produced from filtered LiDAR DEM of El Capitan in Yosemite Valley. Black dashed line delineates the extent of the El Capitan Meadow rock avalanche and black dotted line delineates the distal edge of the active talus slope. Inset shows the major tectonic structures of California, including (in bold lines) the San Andreas Fault (SAF), Owens Valley Fault Zone (OVFZ), White Mountain Fault Zone (WMFZ), and Sierra Nevada Frontal Fault (SNFF). The white square marks the location of Yosemite National Park, and the white star marks the epicenter of the AD 1872 Owens Valley earthquake.

Table 1. Physical parameters for the El Capitan Meadow rock avalanche

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Area of deposit</td>
<td>$2.71 \times 10^5$ m$^2$</td>
</tr>
<tr>
<td>Volume of deposit</td>
<td>$2.19 \times 10^6$ m$^3$</td>
</tr>
<tr>
<td>Volume of intact rock</td>
<td>$1.64 \times 10^6$ m$^3$</td>
</tr>
<tr>
<td>Rock wall source</td>
<td></td>
</tr>
<tr>
<td>Source area</td>
<td>~$8.5 \times 10^4$ m$^2$</td>
</tr>
<tr>
<td>Maximum source elevation (m)</td>
<td>2155</td>
</tr>
<tr>
<td>Minimum source elevation (m)</td>
<td>~1840</td>
</tr>
<tr>
<td>Mean slope of source area (%)</td>
<td>87</td>
</tr>
<tr>
<td>Exposure</td>
<td>Southeast</td>
</tr>
<tr>
<td>Rock type</td>
<td>Granite, granodiorite, minor diorite</td>
</tr>
<tr>
<td>Run out of debris</td>
<td></td>
</tr>
<tr>
<td>Maximum elevation of deposit (m)</td>
<td>1425</td>
</tr>
<tr>
<td>Minimum elevation of deposit (m)</td>
<td>1204</td>
</tr>
<tr>
<td>Maximum vertical drop (m)</td>
<td>950</td>
</tr>
<tr>
<td>Maximum horizontal travel (m)</td>
<td>710</td>
</tr>
<tr>
<td>Energy slope (º)</td>
<td>54</td>
</tr>
<tr>
<td>Fahrböschung (º)</td>
<td>53</td>
</tr>
<tr>
<td>Coefficient of friction (H/L)</td>
<td>1.34</td>
</tr>
<tr>
<td>Total emplacement time (s)</td>
<td>24.1</td>
</tr>
<tr>
<td>Maximum velocity during emplacement (m s$^{-1}$)</td>
<td>100</td>
</tr>
</tbody>
</table>

The deposit has a planimetric surface area of $2.53 \times 10^5$ m$^2$ and a three-dimensional surface area of $2.71 \times 10^5$ m$^3$. The distal portion of the deposit extending beyond the edge of the active talus slope has a mean thickness of 8 m and a maximum thickness of 18 m. The deposit has a hummocky morphology and a clast-supported surface cover of large angular boulders. The distal portion of the deposit extending beyond the active talus slope displays an asymmetric topography along its
length, revealed in a slope-aspect map and in a longitudinal cross-section (Figure 4). Beyond the edge of the active talus, the deposit is relatively fine-grained and has a low topographic gradient, sloping gently upward ~6 m to the crest (Figure 4). The crest is considerably coarser, with boulders several meters in diameter. From the crest, the deposit surface slopes steeply down to the valley floor (Figure 4). This ‘backsloping’ morphology suggests that the deposit is not composite from several rock-fall events. The distal portion is generally very coarse, with large angular boulders projecting several meters above the mean deposit surface (Figure 5A). The largest boulders in the deposit are clustered along the distal margin and are up to ~2700 m³ in volume (Figure 5B).

Determining the exact volume of the rock avalanche deposit is challenging because an undetermined amount of more recent talus has accumulated on top of that portion proximal to the cliff. Because the distal portion of the rock avalanche has a distinct margin and extended out over the nearly horizontal valley floor (Figure 2), the volume for that portion is well constrained by calculating the volume of debris above a horizontal plane at 1204 m amsl (a two-dimensional surface area of 1.03 x 10⁶ m²). This yields a debris volume for the distal portion of 1.03 x 10⁶ m³. This calculation assumes no aggradation of the valley floor over the relatively short time span since the rock avalanche occurred (~3.6 ka; see below). Determining the volume for the proximal portion of the rock avalanche requires an assumption about the extent and thickness of the deposit beneath the active talus. Assuming a tapered wedge beneath the two-dimensional surface area ranging from 1 m thick at the base of the cliff to 8 m where the deposit first appears from beneath the active talus slope, the debris volume for the proximal portion of the rock avalanche deposit is 1.16 x 10⁶ m³. Summing the two portions yields a total deposit volume of 2.19 x 10⁶ m³. This estimate is less than the 3.79 x 10⁶ m³ estimate reported by Wieczorek et al. (1999), probably due to different assumptions regarding the thickness of the rock avalanche deposit beneath the active talus slope, and perhaps also due to the enhanced resolution of the LiDAR-based DEM.

The volume of the rock avalanche deposit can be converted to an approximate volume of the intact rock mass at the source area prior to failure. Assuming a talus density of 1.80 g cm⁻³.
(Sass and Wollny, 2001; Hales and Roering, 2005) and granitic rock density of 2.65 g cm$^{-3}$, the porosity of active talus slope beneath El Capitan should be on the order of 30% (Moore et al., 2009). We observe that the porosity of the rock avalanche deposit is somewhat less than that of the active talus slope, approximately 25%, a value that is in general agreement with other estimates of rock avalanche volume dilation (Nicoletti and Sorriso-Valvo, 1991, and references therein). Reducing the El Capitan Meadow rock avalanche deposit volume by 25% yields an intact rock volume of 1.64 × 10$^6$ m$^3$.

**Rock Avalanche Source Area**

Due to their large volumes, rock avalanches usually have clearly defined source areas, easily identified by large scars or arcuate depressions on hillslopes (Hewitt, 2002; Ballantyne and Stone, 2004; Mitchell et al., 2007). However, El Capitan is unusual in that it does not preserve an immediately obvious source area (Figure 1). The morphology of the rock avalanche deposit indicates that it was sourced from the south-east face, east of the ‘Nose’, the prominent arete at the center of the formation, and west of the ‘North American Wall’, a concave, vertical to slightly overhanging portion of the face named after a mafic dike complex in the shape of North America (Figures 1 and 6D). The width of the rock avalanche deposit constrains the maximum width of the source area to ~550 m; however, because the rock avalanche likely spread laterally across the talus slope and valley floor, the actual source area may have been much narrower.

El Capitan is characterized by a heterogeneous intrusive igneous lithology, which can be used to place further constraints on the rock avalanche source area. Previously published geologic maps show the El Capitan Granite to be the dominant rock type along the south-western flank and at the base of the south-east face, whereas the Taft Granite is prevalent on the south-eastern flank and at the summit (Calkins et al., 1985; Peck, 2002; Ratajeski et al., 2001). There are clear differences in grain size and mafic mineral content between these two lithologies: the El Capitan Granite is a light-colored, coarse-grained biotite granodiorite and granite containing 5–10% mafic minerals, primarily biotite, whereas the Taft Granite is a light-colored, medium-grained, equigranular leucogranite containing 1–5% biotite (Bateman, 1992; Peck, 2002). In addition to these granites, two sets of mafic dikes are exposed on the south-east face, an older, light-colored granodioritic series, and a younger, darker dioritic series whose outcrop pattern resembles the map of North America, and for which the North America Wall is named (Figure 1; Reid et al., 1983; Ratajeski et al., 2001). Numerous thin (<5 m) aplite dikes are also present.

A surficial geologic map (Figure 6A) of the distal portion of the rock avalanche deposit (n = 650 boulders) reveals that the surface is composed of approximately equal amounts of Taft Granite (34%), mafic dikes of the granodiorite series (29%) and El Capitan Granite (27%), with lesser amounts of the North America diorite (9%) and aplite (1%). The various lithologies are broadly distributed and well mixed throughout the distal portion of the deposit, with no clear spatial patterns (Figure 6A). Smaller boulders of El Capitan Granite and diorite become more common with proximity to the active talus slope below El Capitan, reflecting more recent deposition on the active talus after the rock avalanche occurred.

The distribution of rock types in the deposit, in particular the presence of Taft Granite (Figure 6A), can be used to place further constraints on the source area, provided the extent of these rock types on the south-east face is known. Geologic mapping combining field mapping on climbing routes and high-resolution photography (www.xrez.com/yose_proj/Yose_index.html) reveals the outcrop patterns of the various units on the south-east face. Field mapping involved scaled photography of the rock type at each ‘belay’ (anchor point between rope lengths) on several of the major climbing routes ascending the south-east face (n = 72, Figure 6B, 6C). These served as calibration points for more detailed mapping using the high-resolution photographs. Differences in grain size between the El Capitan and Taft granites, as well as differences in the mafic mineral content between these granites and the mafic dikes, allow for accurate delineation of contacts between these units.

A preliminary geologic map of a portion of the south-east face indicates that the Taft Granite is present as a distinct wedge-shaped outcrop below the summit of El Capitan (Figure 6D). A mafic dike of the granodiorite series forms the contact between the Taft and El Capitan granites, and includes small bodies of North America diorite. The El Capitan Granite poses the remainder of the rock within this portion of the south-east face. The outcrop pattern of these rock types, in particular the rather limited extent of Taft Granite only near the summit, constrains the potential source area for the rock avalanche to a maximum elevation of 2155 m and an approximately minimum elevation of ~1800 m (because the El Capitan Granite extends to the base of the cliff, its presence in the rock avalanche deposit does not tightly constrain the minimum elevation of the source area). Within these bounds, the approximate surface area of the rock avalanche source is 8.5 × 10$^4$ m$^2$ (Figure 6D).

Dividing the volume of the intact rock mass (1.64 × 10$^6$ m$^3$) by the surface area of the potential source area, i.e. assuming that the failed rock mass was distributed uniformly across the potential source area, yields a mean slab thickness of 19 m. If the rock avalanche originated from a larger area, the slab thickness would be smaller; conversely, if the rock avalanche originated from only a portion of this area, the slab thickness would be larger. Considering that the rock avalanche source area is presently nearly vertical (dipping ~87° southeast), a rock slab ≥19 m thick would have taken the form of a substantially overhanging rock mass. Although this amount of overhang may seem extreme, the rock mass strength of the granitic rocks composing El Capitan is clearly capable of supporting such structures; the face east of the North America Wall in the vicinity of the climbing route ‘Tangerine Trip’ (Figure 6D) presently overhangs by 37 m horizontally from the base of the cliff. Furthermore, the largest boulder in the distal portion of the rock avalanche deposit (Figure 5B) has a minimum dimension of ~11 m. This represents a minimum thickness for at least part of the rock mass prior to failure (a minimum value because the boulder probably fragmented from a larger rock mass upon impacting the talus slope), supporting the notion of planar or slab failure of a relatively thick rock mass.

**Rock Avalanche Dynamics**

Identifying the potential rock avalanche source area provides insight into the dynamics of the fall and resulting runout. Qualitatively, after detachment the rock mass fell ~650 m down the cliff face, losing potential energy (Figure 7). During the initial fall the rock mass rapidly accelerated due to gravity, with minimal frictional deceleration due to the near vertical cliff angle. The rock mass fragmented upon impact with the apex of the talus slope, causing the various lithologies to...
become well mixed (Figure 6A). Vertical momentum was transferred to horizontal momentum as the debris traveled down the talus slope, dipping south east at a mean angle of 23°, for approximately 395 m to the valley floor (Figure 7). The debris mass was topographically unconfined, with free downward and lateral expansion of the debris across the talus slope and valley floor. As the rock debris moved down the talus slope, the rate of energy dissipation by basal and internal friction became greater than the rate of supply of potential energy. On reaching the flat valley floor the rock mass ceased to lose potential energy, but continued forward due to stored kinetic energy until the last of the kinetic energy was dissipated by friction and the rock mass stopped. This spontaneous stopping led to compression of the distal portion of the rock avalanche deposit, producing the ‘back-sloping’ morphology (Figure 4).

Quantitatively, we determined the maximum velocity and energy balance of the falling rock mass using an algorithm similar to that reported by Wieczorek et al. (2000), which numerically integrates the acceleration due to gravity and deceleration due to friction of a distributed mass in two dimensions (x, y; Figure 7). We modified this algorithm to calculate the average tangential velocity, and also the energy balance in terms of the percentage of potential, kinetic, and...
dissipated energy, as a function of time and tangential distance (Figure 8). We performed the calculation on an elevation profile representing the travel path from the inferred centroid of the source area to the inferred toe of the debris deposit (Figure 7). The apparent coefficient of kinetic friction (or kinetic frictional resistance), $\mu_k = 1.01$, was chosen such that the velocity is zero at the distal edge of the rock avalanche deposit. This coefficient lumps all the mechanisms that act to decelerate the avalanche (e.g. energy dissipation by friction, transformation to thermal and acoustic energy, etc.) into one parameter, and is assumed constant and independent of the tangential velocity, although in practice $\mu_k$ probably decreases with increasing tangential velocity.

The model calculates the tangential velocity, and the partitioning between potential, kinetic, and dissipated energy, as a function of tangential distance and total elapsed time (Figure 8). Assuming instantaneous detachment, the maximum velocity for the rock avalanche calculated by this method is $101 \pm 0.5$ m s$^{-1}$ (safely assumed to be 100 m s$^{-1}$), and it is reached 11.7 s after detachment at a tangential distance of 656 m from the centroid of the source area, where the kinetic energy maxima is also achieved (Figure 8). The total emplacement time is calculated to be 24.1 s (Figure 8). The maximum velocity determined by this method is similar to that determined using an energy slope approach (Hungr et al., 2005; Dortch et al., 2009), in which the maximum vertical distance between the inferred original topography and a line connecting the centroid of the source area with the centroid of the deposit is used to calculate the velocity attained by gravitational acceleration (Figure 7). An energy slope (53°; Figure 7) approach yields a maximum velocity for the El Capitan rock avalanche of 108 m s$^{-1}$.

Rock avalanche dynamics are often evaluated through use of the Fahrböschung (Heim, 1932), the angle of a line connecting the top of the source area to the most distal part of the deposit (Figure 7), and through the ratio of the maximum fall height ($H$) to the maximum runout distance ($L$). These metrics are useful for comparing different rock avalanches, but can greatly overestimate rock avalanche velocities (Hutchinson, 2006). The relatively steep Fahrböschung angle for the El Capitan Meadow rock avalanche (54°; Figure 7) confirms the steep gradient of energy loss and short transport distance. The $H/L$ ratio is 1.34 (Table 1), considerably larger than most reported rock avalanches, which are typically in the range of 0.1–0.6. Energy loss during impact at the base of the cliff substantially reduced the energy available for horizontal motion. Thus, the El Capitan Meadow rock avalanche does not appear to be an example of ‘excessive runout’ landslide, or sturzstrom (Hsu, 1975; Nicoletti and Sorrizo-Valvo, 1991; Davies and McSaveney, 1999). Rather, the deposit extended out onto the floor of Yosemite Valley primarily because of its great potential energy (total potential vertical fall distance of ~950 m), low initial friction, and topographically unconfined travel path.

Figure 7. Profile of the El Capitan Meadow rock avalanche. The rock avalanche source area and extent of deposit are shown in black, inferred where underneath talus. Subsurface bedrock contact from Gutenberg et al. (1956).

Figure 8. Tangential velocity and fraction of potential, kinetic, and dissipated energies as a function of (A) tangential distance of the falling rock mass and (B) total elapsed time. The tangential distance starts at 112 m (the distance from the top of the potential source area to the centroid of the distributed mass). The slope breaks in the velocity, kinetic energy, and dissipated energy correspond to the slope breaks in the rock avalanche profile (Figure 7). The final slope break in the tangential velocity at approximately 1280 m tangential distance corresponds to the maximum thickness of the rock avalanche deposit.
Age of the El Capitan Meadow Rock Avalanche

Determining the age of the El Capitan Meadow rock avalanche is critical for evaluating possible triggering mechanisms, e.g. post-glacial stress changes, meteorological events, or seismic ground motion. Cosmogenic \(^{10}\)Be exposure dating of a prominent recessional moraine and of the rock avalanche itself (Figure 2) constrains the timing of failure.

These deposits are generally well suited for cosmogenic exposure dating because the processes that form them tend to quickly excavate rocks before exposing them on the surface; this is particularly true for rock avalanches, making them an appealing target for exposure dating (Ivy-Ochs et al., 1998, 2009; Barnard et al., 2001; Ballantyne and Stone, 2004; Mitchell et al., 2007; Dortch et al., 2009). Vertical cliffs receive relatively low doses of cosmic rays due to topographic shielding, and the great volumes of rock avalanches suggest that most boulders now exposed on the surface were shielded deep within the cliff prior to failure. Nevertheless, geomorphic deposits must be carefully sampled to ensure reliable exposure dates. Complicating factors can include pre-failure cosmic ray exposure, erosion of the deposit, rotation of boulders after deposition, rock spallation due to forest fires, and topographic, vegetation, and snow shielding of cosmic rays (Bierman and Gillespie, 1991; Gosses and Phillips, 2001; Putkonen and Swanson, 2003; Ivy-Ochs et al., 2009). To address those issues, we collected moraine samples from the tallest boulders on the crest of the moraine, minimizing the effects of erosional degradation (Putkonen and Swanson, 2003). We collected rock avalanche samples from clast-supported, boulder-dominated areas and sampled boulders that were wedged in the deposit and hence could not have rotated since deposition. We minimized the potential for fire-induced rock spallation by sampling only the tops of large boulders. We conducted detailed surveys of the topographic shielding, which is significant beneath the ~1 km tall cliffs (Table II), but determined that snow and vegetation shielding are negligible at this location. We assumed a boulder erosion rate of 0.0006 cm yr\(^{-1}\), a typical value for granitic boulders and bare bedrock surfaces in the Sierra Nevada (Small et al., 1997; Stock et al., 2005).

The glacial history of Yosemite Valley provides a first-order constraint on the age of the El Capitan Meadow rock avalanche: the deposit must post-date glacial occupation of the valley below El Capitan, for any rock debris falling onto the glacier would have been transported to the terminal moraine several kilometers down the valley. Cosmogenic \(^{10}\)Be concentrations in three large boulders on the crest of the El Capitan recessional moraine, located 850 m down valley from the rock avalanche deposit (Figure 2), yield exposure ages between 16.4 and 23.5 ka, with a mean age of 19.8 ka (Table III). This age is consistent with previous estimates for the Tioga glaciation (Philips et al., 1996; 2009; Clark et al., 2003). LGM deglaciation likely occurred rapidly in the Yosemite region (Clark et al., 2003), with the area around El Capitan probably becoming ice-free by ~15–17 ka. Thus, the El Capitan Meadow rock avalanche must be younger than ~15–17 ka.

Four of the five sampled boulders on the rock avalanche deposit yield \(^{10}\)Be exposure ages clustered between 3.50 and 3.84 ka, within analytical uncertainty, and yield an error weighted mean age of 3.6 ± 0.2 ka (Table II). These samples span the width of the deposit, indicating that it either resulted from a single event or from multiple events relatively closely spaced in time (i.e. within the analytical uncertainty in the data). A fifth boulder yields a much older age of 21.9 ka (Table II). This age represents a clear outlier from the other samples, and is at odds with the deglaciation history described above. Exposure ages from the El Capitan recessional moraine suggest that it is unlikely that even a portion of the rock avalanche occurred at 21.9 ka because the LGM glacier probably still occupied the flank of El Capitan at that time. The 21.9 ka exposure age does not appear to represent the timing of failure, and instead results from inherited nuclide concentration due to exposure on the cliff prior to occurrence of the rock avalanche (Ivy-Ochs et al., 2009), i.e. the sampled boulder surface was at or near the cliff face prior to failure. This result confirms the need to collect and analyze multiple samples from each deposit to be dated (Putkonen and Swanson, 2003), even for deposits, such as rock avalanches, that have relatively simple depositional histories. Excluding the 21.9 ka outlier age yields an otherwise consistent age of ~3.6 ka for the El Capitan Meadow rock avalanche.

Destabilizing Factors and Possible Triggering Mechanism(s)

Post-glacial adjustment of rock slopes reflects the interaction of changing stress conditions, resulting from glacial oversteepening and relaxation of residual stresses following debuddressing, and rock mass strength, controlled by lithology and discontinuities (Ballantyne, 2002). Studies of rock slopes in formerly glaciated areas suggest that oversteepened slopes

### Table II. Cosmogenic beryllium-10 data and exposure ages for El Capitan moraine and rock avalanche boulders

<table>
<thead>
<tr>
<th>Sample</th>
<th>Deposit type</th>
<th>Latitude</th>
<th>Longitude</th>
<th>Elevation (m amsl)</th>
<th>Mass qz (g)</th>
<th>(^{10})Be/Be (10^(^{-14}))</th>
<th>Shielding</th>
<th>(P_{0}) (atm g (^{-1}) yr(^{-1}))</th>
<th>(^{10})Be (10^(^{19}) atm g (^{-1})qtz)</th>
<th>Exposure age (ka)</th>
</tr>
</thead>
<tbody>
<tr>
<td>ECM-1</td>
<td>Moraine</td>
<td>37.7228081</td>
<td>−119.643703</td>
<td>1215</td>
<td>20−25</td>
<td>237.40 ± 33.52</td>
<td>0.8844</td>
<td>9.81</td>
<td>22.69 ± 3.37</td>
<td>23.53 ± 4.59</td>
</tr>
<tr>
<td>ECM-2</td>
<td>Moraine</td>
<td>37.7227539</td>
<td>−119.643701</td>
<td>1216</td>
<td>55−70</td>
<td>535.00 ± 18.68</td>
<td>0.8844</td>
<td>9.74</td>
<td>18.92 ± 6.83</td>
<td>19.37 ± 2.02</td>
</tr>
<tr>
<td>ECM-3</td>
<td>Moraine</td>
<td>37.7231261</td>
<td>−119.643987</td>
<td>1216</td>
<td>43−37</td>
<td>359.30 ± 27.63</td>
<td>0.8844</td>
<td>9.58</td>
<td>16.02 ± 1.28</td>
<td>16.43 ± 2.11</td>
</tr>
<tr>
<td>ECL-1</td>
<td>Rock avalanche</td>
<td>37.7253064</td>
<td>−119.631339</td>
<td>1219</td>
<td>36−19</td>
<td>69.31 ± 7.93</td>
<td>0.8609</td>
<td>9.39</td>
<td>3.84 ± 0.59</td>
<td>3.78 ± 0.68</td>
</tr>
<tr>
<td>ECL-3</td>
<td>Rock avalanche</td>
<td>37.7254811</td>
<td>−119.632579</td>
<td>1221</td>
<td>38−08</td>
<td>66.25 ± 0.71</td>
<td>0.8557</td>
<td>9.38</td>
<td>3.55 ± 0.32</td>
<td>3.49 ± 0.45</td>
</tr>
<tr>
<td>ECL-4</td>
<td>Rock avalanche</td>
<td>37.7260609</td>
<td>−119.628353</td>
<td>1211</td>
<td>36−09</td>
<td>69.62 ± 0±03</td>
<td>0.8622</td>
<td>9.38</td>
<td>3.83 ± 0.47</td>
<td>3.77 ± 0.58</td>
</tr>
<tr>
<td>ECL-6</td>
<td>Rock avalanche</td>
<td>37.7260881</td>
<td>−119.610599</td>
<td>1214</td>
<td>42−30</td>
<td>147.80 ± 46.57</td>
<td>0.9457</td>
<td>9.27</td>
<td>20.15 ± 2.82</td>
<td>21.94 ± 4.06</td>
</tr>
<tr>
<td>ECL-8</td>
<td>Rock avalanche</td>
<td>37.7252681</td>
<td>−119.644443</td>
<td>1211</td>
<td>34−05</td>
<td>60.66 ± 7.27</td>
<td>0.8609</td>
<td>9.33</td>
<td>3.50 ± 0.59</td>
<td>3.46 ± 0.67</td>
</tr>
</tbody>
</table>

\(^{1}\)Be/Be ratios normalized to NIST standards (Nishizumi et al., 2007). Uncertainty is 1σ analytical uncertainty.

\(^{2}\)Local \(^{10}\)Be production rate, incorporating spallogenic and muogenic production scaled for latitude, altitude, and topographic shielding (Balco et al., 2008).

\(^{3}\)Exposure ages calculated using the CRONUS calculator (Balco et al., 2008) and assuming a \(^{10}\)Be half-life of 1.36 ± 0.07 Myr (Nishizumi et al., 2007), an attenuation length scale of 160 g cm\(^{-2}\), rock density of 2.65 g cm\(^{-3}\), and the spallation scaling scheme of Lal (1991) and Stone (2000).

tend to adjust rapidly to these stress changes by rock mass failure, but that the timing and magnitude of these failures are strongly controlled by rock mass strength (Augustinus 1995a,b). The vertical to overhanging face of El Capitan has clearly been steepened by glacial erosion, but the timing of the rock avalanche and the approximate location of the source area suggest that it did not fail in response to stress release associated with deglaciation. Although rock mass failures related to debretressing may lag deglaciation by centuries or even millennia (Ballantyne, 2002; Ballantyne and Stone, 2004), the ~13 kyr lag between deglaciation of Yosemite Valley and failure of the El Capitan Meadow rock avalanche suggests other triggering mechanisms. As described above, the El Capitan Meadow rock avalanche appears to have originated from near the summit of El Capitan, which is well above the LGM limit and has probably not been glaciated in at least 800,000 years (Huber, 1987). Large rock falls tend to be more common from the upper slopes of Yosemite Valley, above the limit of LGM glaciation, where rocks tend to be more weathered (Wieczorek et al., 1999; Wieczorek, 2002). However, we note that the rocks composing the upper south-east face of El Capitan are only slightly weathered. Intersections between surface parallel sheeting joints and vertically oriented discontinuities, which are more prevalent in the Taft Granite, likely formed zones of weakness along which detachment occurred. In addition, lithologic contacts and dikes within the source area (Figure 6D) may have helped to destabilize the rock mass; for example, Weidinger and Korup (2009) determined that leucogranite dikes in the source area contributed to a large rock slope failure in the Himalaya.

Triggering of large rock slope failures is often attributed to thresholds associated with either intense precipitation events (Wieczorek, 2002; Malamud et al., 2004; Gabet et al., 2004) or large magnitude earthquakes (Keefee, 1984, 1994; Barnard et al., 2001; Wieczorek, 2002; Malamud et al., 2004; Strom and Korup, 2006; Owen et al., 2008; Yin et al., 2009) note, however, that rock avalanches have also been shown to occur in the absence of any obvious triggering mechanisms (McCave, 2002). It is difficult to evaluate the potential role of an extreme precipitation event in triggering the El Capitan Meadow rock avalanche because of uncertainty in both the exposure dating of the deposit and in the chronology of local paleoclimate records. Moreover, intense but short duration precipitation events may not be preserved well in the geologic record. Existing geologic and paleoclimate data do not record such an event, nor do they suggest an abnormally wet climate for an event of this magnitude. Existing geologic and paleoclimate data do not record precipitation events may not be preserved well in the geologic record. Existing geologic and paleoclimate data do not record an extreme precipitation event in triggering the El Capitan rock avalanche; indeed, such precipitation events are known to have triggered historic large rock slope failures in the Sierra Nevada (Wieczorek, 2002). Furthermore, many large rock falls and rock slides in Yosemite have occurred in the absence of recognized triggering mechanisms (e.g. the 10 March 1987 Middle Brother rock fall of ~600,000 m3; Wieczorek and Snyder, 2004). However, the very large volume of the El Capitan Meadow rock avalanche deposit, the convergence of exposure dates with the estimated timing of the penultimate rupture of the Owens Valley fault, and the fact that the AD 1872 Owens Valley fault earthquake generated very large rock falls in Yosemite Valley, all support a seismic trigger. Mathes (1930) may well have been correct in describing the El Capitan Meadow rock avalanche deposit as an ‘earthquake avalanche’.

There are at least five other large rock fall deposits in Yosemite Valley that have been interpreted as rock avalanches, including one near Mirror Lake in eastern Yosemite Valley with a much larger volume that of the El Capitan Meadow rock avalanche (~1.1-4 × 106 m3; Wieczorek et al., 1999; Wieczorek, 2002). Given the large size of these deposits, it is plausible that they too were triggered by earthquakes, and perhaps by the ~3.6 ka Owens Valley and/or White Mountain faulting event. Additional 10Be exposure dating of these deposits will help to test this hypothesis. If these rock avalanches are also found to coincide with seismic events, it would suggest that the extreme hazard posed by rock avalanches in Yosemite Valley remains present and depends on local earthquake recurrence intervals, emphasizing the need for detailed paleoseismic data from the Sierra Nevada frontal fault system and adjacent faults.

Conclusions

The morphology and age of the 2.19 × 106 m3 El Capitan Meadow rock avalanche deposit in Yosemite Valley indicates a massive failure of a large expanse of the south-eastern face of El Capitan. Geologic mapping of the deposit and of the south-east face constrains the source of the rock avalanche to an area near the summit of ~8.5 × 106 m3. The rock mass free fall ~650 m, attaining a maximum velocity of 100 m s^-1 before impacting the talus slope and spreading across the valley floor. Four cosmogenic beryllium-10 exposure ages from boulders in the deposit are tightly clustered between 3.5 and 3.8 ka, with a mean age of 3.6 ± 0.2 ka. These samples span the width of the deposit, confirming that it resulted from a single event.
A fifth boulder gives a much older age of 21.9 ka, at odds with the glacial history of Yosemite Valley and likely resulting from inherited nuclide concentrations close to exposure on the cliff prior to failure. Because the rock avalanche occurred well after deglaciation, glacial debuddling is an unlikely trigger. However, the age of the El Capitan Meadow rock avalanche coincides with estimated rupture of the Owens Valley fault between 3.3 and 3.8 ka. This coincidence of ages, combined with the fact that the AD 1872 Owens Valley earthquake triggered numerous large rock falls in Yosemite Valley, suggests that a large magnitude (≥M7.0) earthquake centered in the south-eastern Sierra Nevada ca 3.6 ka may have triggered the El Capitan Meadow rock avalanche. If correct, this suggests that the hazard posed by rock avalanches in Yosemite Valley remains present and depends on local earthquake recurrence intervals.

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References


