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Glacial erosion and geomorphology in the northwest Sierra Nevada, CA

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Abstract

Pleistocene glacial erosion left a strong topographic imprint in the northwestern Sierra Nevada at many scales, yet the specific landforms and the processes that created them have not been previously documented in the region. In contrast, glaciation in the southern and central Sierra was extensively studied and by the end of the 19th century was among the best understood examples of alpine glaciation outside of the European Alps. This study describes glacially eroded features in the northwest Sierra and presents inferred linkages between erosional forms and Pleistocene glacial processes. Many relationships corroborate theoretical geomorphic principles. These include the occurrence of whalebacks in deep ice positions, roches moutonnées under thin ice, and occurrence of P-forms in low topographic positions where high subglacial meltwater pressures were likely. Some of the landforms described here have not previously been noted in the Sierra, including a large crag and tail eroded by shallow ice and erosional benches high on valley walls thought to be cut by ice-marginal channels.

Keywords: Quaternary landforms; Erosion; Alpine glaciation

1. Introduction: knowledge of Sierra glaciation

This paper describes Pleistocene glacial landforms in the NW Sierra Nevada, CA and links them with glacial processes. Erosional landforms dominate the landscape and are emphasized because of their ubiquity and persistence across the severely eroded landscape. Considerable study of glaciation in the southern and central Sierra over the last 140 years provided an early understanding of Sierra glacial geomorphology and stratigraphy and, to some extent, basic principles drawn from those studies are directly applicable to the NW Sierra Nevada. In some ways, however, the geologic structure and history of the northern Sierra Nevada are very different and variations in the geomorphology reflect the differences. Much has been learned about glaciological processes over the last 50 years. This knowledge has not been applied to glacial features in the Sierra Nevada because paleoglacial research there has moved away from linking landforms and glaciological processes to an emphasis on stratigraphic questions. Early researchers (e.g., Gilbert, 1904) were fascinated by glacial landforms to the south, but scientific understanding of subglacial processes was limited. Few modern studies have been concerned with glacially eroded landforms in the Sierra Nevada.

This study presents a survey of geomorphic features ranging from small, local striae and gouges to large

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glacial troughs, roches moutonnées, and a crag and tail. While qualitative in nature, these geomorphic descriptions provide indicators of important glacial processes and constraints on their former patterns and characteristics. The intent of this paper is to (i) describe glacially eroded forms in the NW Sierra Nevada and (ii) link these forms to modern concepts of glacial processes in order to facilitate inferences about the nature of glaciation at the time the landforms were generated. The glaciological literature is far too extensive and complex to be covered in this brief treatment, so discussion is restricted to a few key examples. Hopefully, future geomorphic studies will advance the linking of form to process and will integrate glaciological principles elsewhere across the Sierra.

1.1. Early application of glacial theory in California

Introduction of the glacial theory to formal western science was followed rapidly by the arrival of European culture in the California interior. The radical new geologic ideas about glaciation were transported from Europe across the English Channel and Atlantic Ocean, then across North America to California with astonishing speed. Evidence of alpine glaciation in the mountains of California was very quickly recognized and soon became the best-known example of alpine glaciation in the New World. Much as the settlement of California leaped ahead of settlement on the western frontier, so application of the glacial theory bounded across the western interior into California.

The gathering of geologists into the California Geological Survey in the early 1860s brought Josiah Whitney and William Brewer fresh from geological training at Yale and travels in the European Alps, together with Clarence King fresh from course work under Louis Agassiz at Harvard (Guyton, 1998). Although Whitney (1865) downplayed the geomorphic importance of glacial erosion, the discovery and documentation of extensive glacial evidence in the central Sierra Nevada in the early 1860s provided examples of the wide application of the new theory (Whitney, 1865; King, 1872; Brewer, 1966). Joseph LeConte, another student of Agassiz, arrived in California from South Carolina in 1869, took a post at Berkeley, and soon introduced several important glacial concepts to the educated public, John Muir, and the scientific community. He traveled through the south Tahoe basin and down the South Fork American River and provided the only nineteenth century scientific descriptions of glacial evidence in the northern Sierra (LeConte, 1873) until Lindgren (1897, 1900) recognized glacial deposits in mapping the general geology of the area. John Muir's accounts of the Yosemite Valley spread glacial knowledge to a wide audience because of the novelty of the glacial premise and the grandeur and passion of his landscape descriptions (e.g., Muir, 1873a,b).

1.2. The next generations of glacial geomorphology

Fascination for alpine glacial geomorphology in the Sierra continued through the turn of the century when relationships between form and process at all scales received scrutiny that has not since been paralleled in the region. G.K. Gilbert, who had written earlier of the pre-glacial Sierra topography (Gilbert, 1883), revisited the central Sierra to study small- and intermediate-scale glacial features and processes (Gilbert, 1904, 1906a,b). Johnson (1904) concluded that glacial erosion had created and deepened valleys by back-wearing of cirques.

As the concept of multiple glaciations arose, attention began to shift from geomorphic process–form relationships to the complexities of glacial stratigraphy. The work of Russell (1889) in the east–central Sierra had clearly documented at least two major glacial advances and several lesser "fluctuations." He noted multiple lateral moraines and recognized a gravel deposit separating lacustrine beds in Mono Lake as evidence for a period of interglacial lake dessication. Ultimately, glacial studies in the southern and central Sierra Nevada by Matthes (1930) and Blackwelder (1931) provided a basic stratigraphic framework for Pleistocene glaciations. Matthes (1930) was deeply concerned with geomorphic questions around Yosemite. He extrapolated valley-side profiles across canyon inner gorges to illustrate depths of canyon incision and defined three erosional surfaces in the Yosemite area. Matthes (1930) also recognized three glacial advances in the Yosemite area and qualitatively noted general characteristics of the moraine morphologies, boulder frequencies, and weathering characteristics. Blackwelder (1931) recognized multiple glacial advances east of Yosemite Valley across the Sierra crest and correlated them with Matthes' (1930) mapping units.

1.3. Glacial knowledge in the NW Sierra

In spite of early recognition and on-going study of glaciation in the southern and central Sierra Nevada, little work was done in the NW Sierra where the extent of ice remained unknown until the turn of the twentieth century. LeConte (1872, 1873) briefly visited and described glacial evidence in the area south of Lake Tahoe and down the South Fork American River. Yet, this information was not well known. Chamberlin's (1888) map of the extent of ice in the United States showed no glacial ice in the Sierra Nevada north of Carson Pass or the Stanislaus River (Fig. 1).

The dearth of glacial geomorphic knowledge of the NW Sierra persisted through most of the 20th century. Several geologic maps of the region included Pleistocene deposits (Lindgren, 1897, 1900; Hudson, 1951; Harwood, 1980), but they were primarily interested in older rocks and did not identify glacial landforms. Jones (1929) provided general descriptions of the geography of glaciated areas south of Lake Tahoe

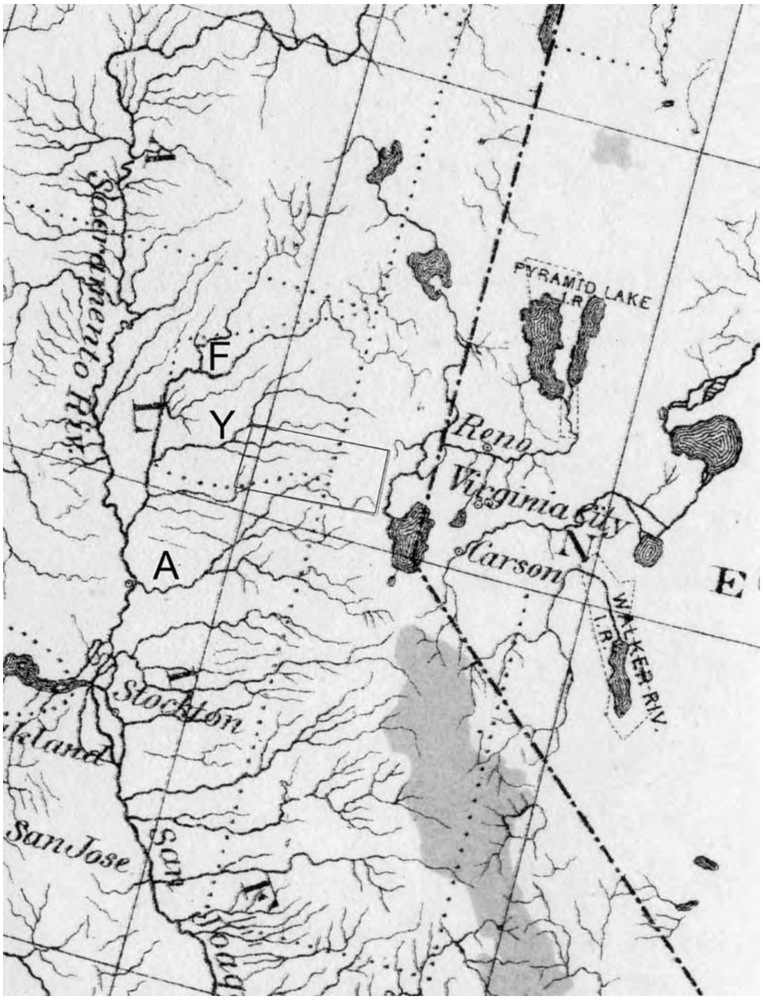

Fig. 1. Detail of northern California from Chamberlain's (1888) map of glaciation (shaded) in the United States. The northern glacial limit terminates south of Lake Tahoe indicating no Pleistocene glaciation of NW Sierra. Apparently, Chamberlin was not aware of LeConte's (1873) surveys. Box (added) shows study area of this paper. Letters (added) represent American (A), Yuba (Y), and Feather (F) Rivers.

and Desolation Valley in the South Fork American drainage, but did not consider processes. Blackwelder (1931) did not map in the NW Sierra, but he described three stratigraphic units along the Southern Pacific Railroad. South of Bear Valley near the Blue Canyon Station he interpreted two old till outcrops as Sherwin and extensive upland till deposits as Tahoe. He described Tahoe ice as 180 m thick at Cisco Station and 3 km wide and more than 300 m thick where it capped the ridge above Emigrant Gap. Blackwelder also described small Tioga moraines near the Soda Springs railroad station as having "weak cirques,... and some very low moraine loops at an altitude of 6500 feet [1980 m] a few miles farther west. The glacier was only about 200 feet [656 m] thick and 5 miles [8 km] long" (Blackwelder, 1931, p. 909). Recent mapping and surface-exposure ages corroborate high pre-Tioga (younger Tahoe?) tills at Emigrant Gap but show that Tioga ice was at least 320 m thick above the valley bottom at Cisco and extended through Bear Valley, much thicker and more extensive than Blackwelder's estimate (James et al., 2002).

In the 1960s, reviews of the Quaternary geology of the Sierra range included small-scale maps showing Tioga glaciers in the northern Sierra Nevada (Wahrhaftig and Birman, 1965; Bateman and Wahrhaftig, 1966). In the NE Sierra, Birkeland (1964) mapped deposits and developed a glacial stratigraphy around Donner Lake and the Truckee River. He applied Blackwelder's (1931) Tahoe and Tioga units to the dominant moraines around Truckee, interpreted the Tahoe as Wisconsin in age, and recognized two pre-Wisconsin tills he named Donner Lake and Hobart Mills. He also mapped a post-Tioga Frog Lake till confined to high cirques such as on the east flank of Castle Peak. No other large-scale glacial stratigraphic or geomorphic mapping had been done until recently (James and Davis, 1994; James, 1995; James et al., 2002).

At least three major Pleistocene glacial advances occurred in the region and each of them may have had multiple stades (James et al., 2002). While small-scale geomorphic features are likely to be products of the latest glaciation to override them, larger landforms may be polygenetic. They were not only eroded by multiple glaciations but also experienced substantial weathering intervals during interstadials. The following discussions of glacial ice concentrate on the nature of the Tioga glaciers at their utmost extent, i.e., the last glacial maximum, although at least two older glaciations reached slightly higher elevations.

2. Physiography of the study area

2.1. Geography

The NW Sierra Nevada includes the drainages of the three forks of each of the American, Yuba, and Feather Rivers. These rivers flow west from the Sierra crest (Fig. 1) with the South Fork American beginning SW of Lake Tahoe and the northernmost North Fork Feather River slightly north of the latitude of Pyramid Lake in Nevada. The study area is primarily in the South Yuba drainage, although ice flowed out into the Middle Fork Yuba, the North Fork American, and Bear Rivers (Fig. 2). The geology of this region is similar to the central and southern Sierra in that much of the higher elevations are underlain by Mesozoic granitic batholiths that have been overridden by ice, and at lower elevations deformed Paleozoic metamorphics are common with ridges capped by flat-lying Cenozoic volcanics. However, rock types are more diverse at high elevations of the NW Sierra than to the south where granite predominates. Both metamorphic basement rocks and a Cenozoic cover of conglomerates and volcanics are more widespread here. The topography is also more subdued in this region, with the exception of the North Fork American River along the south margin of the study area.

2.2. The preglacial landscape

Preservation of an extensive Tertiary channel system in the northern Sierra provides important constraints on landform evolution in the region (Lindgren, 1911). Eocene channels were incised into a landscape that has been interpreted as a broad eroded upland with rolling hills and tropical weathering (Yeend, 1974). For example, along the lower South Yuba River, the main ancestral Yuba channel had cut ~ 300 m below a surface defined by the tops of San Juan, Washington, and Harmony Ridges. The present elevation of modern channels below the Tertiary channels represents canyon deepening that was encouraged by uplift of the Sierra block (Lindgren, 1911).

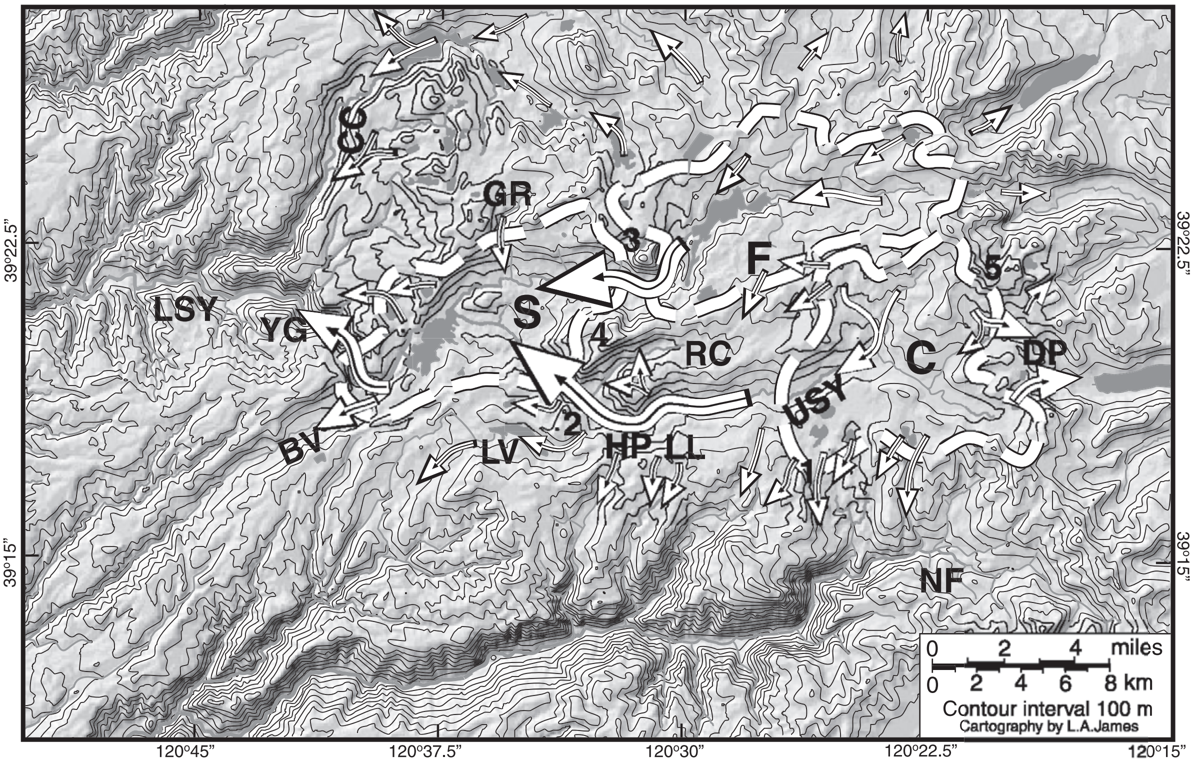

Fig. 2. Tioga glacial flows (arrows) and ice fields (dashed lines). Topography from Chico-E 1° DEM (1:250,000 base). C = Castle ice field, F = Fordyce ice field, S = Spaulding ice field. 1 = Devils Peak, 2 = two roches moutonnées: Cisco Butte and Hill 6642, 3 = Old Man Mountain, 4 = Red Mountain, 5 = Castle Peak. BV = Bear Valley, CC = Canyon Crk., DP = Donner Pass, GR = Grouse Ridge, HP = Huysink Pass, LL = Loch Leven Pass, LSY = lower So. Yuba Cn., LV = Lake Valley, NF = No. Fk. American Cn., RC = Rattlesnake Cn., USY = upper So. Yuba Cn., YG = So. Yuba Gorge.

During a late Cenozoic volcanic period, the preexisting drainage filled with sediment and a series of new valleys developed flowing WSW. These valleys formed parallel drainage patterns as is common with over-steepened channel systems (Howard, 1967), presumably in response to uplift and steepening of the Sierra block. Deep channel incision ensued so that the modern drainage is much lower than the pre-volcanic Tertiary channel system. For example, the modern lower South Yuba Canyon is ∼250 m below the level of the Eocene channel bed near the town of Washington. Erosion below the ancestral channel probably was not substantial until the voluminous production of volcaniclastic rock began to subside in the Pliocene. Thus, much incision in the Yuba and American Canyons apparently occurred during the Quaternary.

Hanging valleys (tributaries that are discordant to main valleys) were first named by G.K. Gilbert while inspecting Alaskan fjords (letter to Wm. M. Davis in 1899; Davis, 1900). Although this connection between glacial erosion and hanging valleys is often assumed, hanging valleys are quite common in the Sierra Nevada foothills below the glacial limit. Thus, unfortunately, neither the degree of glacial erosion nor the occupation of valleys by a glacier can be established simply from the presence of hanging valleys in this region. Discordant tributaries in the lower South Yuba and North Fork American Canyons are the result of steepening of the main E–W trending valleys by late Cenozoic uplift that left tributaries relatively unaffected (Lindgren, 1911; Matthes, 1930). The increased fluvial erosive power and incision rates of main channels left tributaries hanging. Stream piracy of the former upper Bear River also played a role in the South Yuba by increasing stream power and incision rates (James, 1995).

At higher elevations in the glaciated valleys of the South Yuba, hanging valleys are best expressed in the South Yuba gorge below Bear Valley (Fig. 2) (e.g., James, 1995, Fig. 8). Hanging tributaries here could not have been formed prior to the Pliocene because a large Tertiary channel passed through the ridge at Bear Valley only 1 km to the south of the gorge. This paleochannel is filled with andesitic lahar material presumably deposited between the late Miocene and early Pliocene (Slemmons, 1960). Ironically, at higher glaciated elevations hanging valleys are lacking along the South Yuba or Fordyce Canyons and modern streams tend to be graded to valley bottoms.

2.3. Ice fields and valley glaciers

Tioga (Late Wisconsin) glaciers formed a series of ice fields and valley glaciers that not only flowed down into the lower study area but also spilled across drainage divides in all directions. The following description of glaciers at Tioga maxima is based on topographic relationships and, in some locations, field mapping of striae and other directional features. The glaciological terminology used here (Table 1) is conventional for glaciers constrained by topography with one exception: small, steeply descending glaciers issuing from an upland ice field are referred to as outlet glaciers, a term usually applied to steep glaciers flowing out of large ice sheets or ice caps. Most of the glaciers described here were connected to some degree, so the glacier system could be broadly described as one large ice field at high elevations feeding down to an extensive transection glacier. This general description can be subdivided, however, into two high ice fields feeding a lower ice field through two major valley glaciers. The two upper ice fields extended across the Sierra crest steeply down to the east (see Birkeland, 1964), although descriptions here stop at the crest. They also formed a more or less continuous band in the N–S direction along the crest, although they are separated in the following discussion based on the canyons they fed to the west.

Table 1
Types of glaciers constrained by topography[a]

Icefield	Large expansive ice masses free to flow so ice doming is absent.
Valley glacier	Ice flowing in a deep bedrock valley; may include branching system.
Transection glacier	Interconnected valley glaciers in a web-like pattern with minor flows between valleys.
Cirque glacier	Ice emanating from small basins in valley heads.
Outlet glacier	Ice flowing steeply out of a high accumulation area; may be associated with rapid flow, crevasses, or ogives.
Small glacier	Niche glaciers, glacierets, ice aprons, ice fringes.

[a] Sugden and John (1976) and Benn and Evans (1998).

Three broad ice fields occupied large topographic depressions: the Fordyce, Spaulding, and Castle ice fields (Fig. 2). The Fordyce ice field extended ∼ 8 km north to south and 13 km from the Sierra crest west to Old Man Mountain. Some Fordyce ice spilled north into the Middle Yuba Canyon, some flowed south to the South Yuba through Rattlesnake Canyon, but most flow was to the west in a valley glacier ∼ 600 m thick between Old Man and Red Mountains. The Fordyce valley glacier flowed a few kilometers down to a lower ice field around Lake Spaulding, which was ∼ 9 km long by 5 km north to south. An outlet glacier entered the Spaulding ice field from the accumulation area north of Grouse Ridge. It formed a lateral moraine along Granite Creek that descended steeply 130 m in elevation ending abruptly at the level of the Spaulding ice field surface. Ice from the Spaulding ice field discharged through three valley glaciers. The main discharge flowed steeply down the deep South Yuba Gorge. A broad valley glacier more than 200 m thick extended SW through Bear Valley, although its low gradient and a constriction at the base of Bear Valley moderated flow. A broad outlet glacier flowed west across a high shallow trough NW of Lake Spaulding and spilled into the South Yuba Gorge as an ice fall.

To the east near the Sierra crest at Donner Pass, the Castle ice field extended across a basin 8 km wide and more than 10 km north to south. This ice field covered both Upper and Lower Castle Valleys that were connected across an arête at Castle Gate, although little ice flowed between these upper accumulation areas. Some ice flowed between Castle and Fordyce ice fields, but flows were primarily to the east, south, and west. To the south, ice spilled from the Castle ice field in a broad thin sheet across a plateau into the deep North Fork American Canyon and, to the west, ice fed a deep valley glacier in South Yuba Canyon. Ice from Upper Castle Valley bifurcated and flowed both east and west. Birkeland (1964) postulated correctly that the Tioga ice divide was west of the topographic divide, although the ice divide was not more than ~ 2 km west of Donner Pass.

Most flow from the Castle ice field was west down South Yuba Canyon where it fed a valley glacier more than 300 m deep. Much of the valley glacier in this part of the canyon topped a low plateau ~ 200 m above the South Yuba valley bottom at several points along the southern divide and flowed into the American River drainage in relatively thin sheets on the order of 100 m thick. The ice-surface slope to the west was steeper than the ridge line of the southern divide, so the ridge emerged from the ice to the west. Consequently, ice flows south into the North Fork American Canyon decreased westward along the South Yuba valley glacier. Yet, even in the lower west end of this canyon, some ice spilled south to the North Fork American Canyon through high passes in the Loch Leven and Huysink Lake areas. Near the mouth of the upper South Yuba Canyon, ice topped Cisco Butte and Hill 6640, two roches moutonnées that impeded flow. The dominant flow of the South Yuba valley glacier passed NW through a deep trough between Cisco Butte and Red Mountain to the Spaulding ice field. Substantial discharges of ice also spilled into Lake Valley and behind Cisco Butte. These glaciers ultimately flowed SW to the North Fork American drainage.

3. Glaciation and glacially eroded landforms

This discussion focuses on erosional landforms that predominate in the region. Till deposits and lateral moraines are common in the area (James, 1995), but they are far less extensive than the erosional features. Glacial erosion is generally driven by four processes: abrasion, plucking, subglacial meltwater, and dissolution (Benn and Evans, 1998). Abrasion is greatest on the stoss sides of protruding rock masses, is reliant on basal debris, and is discussed under the sections on polish, striae, grooves, and gouges. Plucking is greatest on the lee side of protruding rock masses and is discussed under the section on roches moutonnées. Subglacial meltwater can be important where high pressures develop and is discussed in the sections on P-forms and roches moutonnées. Dissolution can be an important mode of glacial erosion in calcareous rocks, but these are not common in the area and this mode of erosion is not discussed.

3.1. Polish, striae, and grooves

Interactions between sediment in the basal layer of temperate glaciers and the bedrock surface may create small-scale erosional landforms such as polish, striations, or grooves. These abrasion features are common in the NW Sierra, although their spatial distribution varies with preservation and paleoenvironments. Preservation of Pleistocene polish and striae is favored by hard rock that resists weathering and by surface conditions that have been protected from weathering and erosion. The ubiquitous granodiorites in the area are prone to exfoliation and disintegration and do not preserve surface features well, although polish and striae are common on Tioga-age granodiorite surfaces (Fig. 3). No known example of striae on a pre-Tioga surface has been found in this area, although Matthes (1930, p. 70) describes one such location near artist's Point in the Yosemite region as an anomaly. Abundant polish, striae, and grooves can be found on metamorphic rocks, particularly at the north end of Bear Valley and the entrance to South Yuba Gorge.

Patterns of basal erosion result from variations in basal pressures, ice velocity, and sediment (Boulton, 1979) as well as rock structures. Processes that favor abrasion include basal melting and sliding with abundant basal sediment that is hard relative to the bed. Basal melting causes a downward ice velocity that forces basal clasts against the bed (Iverson, 1991a). Basal melting is common on stoss sides of protruding rock masses and under thick ice. Conditions that promote rapid sliding include steep gradients, a wet bed with ice near the pressure–melting point, and high hydrostatic pressures, although the latter condition may reduce abrasion. The abundance of basal sediment is affected by local bed erosion and the introduction of sediment from glacier margins. Without sediment, the ability of the ice to abrade its bed is severely limited. High shear stresses are not necessary for abrasion, so polished and striated surfaces can occur under relatively thin ice.

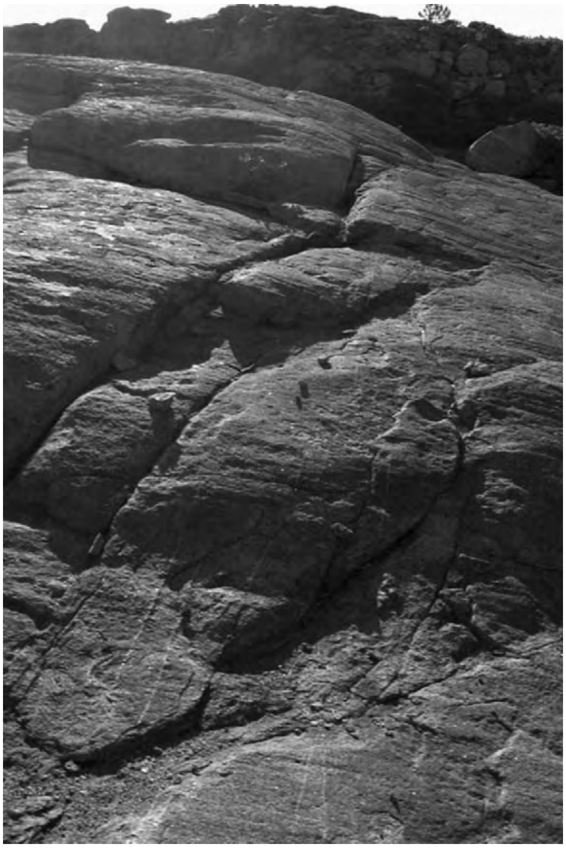

Fig. 3. Striae and polish on granodiorite in Loch Leven provide evidence of Tioga ice flowing south across a high pass. Ice flowed right to left.

Glacial polish is a highly smoothed rock surface caused by abrasion. Polish represents the work of fine-grained sediment as opposed to striae that represent the focused translation of energy from large clasts to the bedrock. Most glacially smoothed surfaces have a much greater area of polished surface than area of striae and grooves. This led Hindmarsh (1996) to conclude that erosion by fine-grained sediment far surpasses erosion by larger particles. This concept is supported by the fact that large clasts break down to fine-grained sediment that continues to polish. True polish with a high sheen is found only in isolated patches on the Pleistocene glacial surfaces in the study area. Polished surfaces are common, however, with a smooth but granular surface due to post-glacial weathering of selected surface mineral grains to a depth of less than 1 mm. Slightly weathered polish is common on fine-grained metamorphic rocks, quartz veins, and on granodiorites where exfoliation has not yet occurred. Extensive polished surfaces in the area suggest the former presence of fine-textured basal till that has largely been stripped away.

Glacial striae are linear scratches caused by dragging coarse clasts along the basal contact of the glacier. Three forms of striae, noted by Chamberlin (1888), were related by Iverson (1991a) to striater point sharpness and rotation. Iverson concluded that striation depth was a function of the propensity for a clast to rotate and that nonspheroidal angular fragments produce the deepest striae because they resist rotation. Nailhead striae (Fig. 4), a variant of Iverson's (1991a) Type I striae, can serve as directional indicators because they result from shifting of the clast at the down-ice limit of the striation (Benn and Evans, 1998). Striae are common throughout the region, most frequently on stoss sides of rock protrusions but also on flat surfaces and along valley walls. They can be found in locations that were under thick and thin ice. They are best preserved on hard metamorphic rocks but can be found on some granodiorites where weathering has been limited.

Glacial grooves are deep linear depressions representing abrasion concentrated along an ice flow path. They are oriented parallel to ice-flow directions and often occur in massive rocks, so they are not generally the result of structural weaknesses in the rock such as bedding planes or folia. Boulton (1979) described debris-rich bands of ice above grooves under the Breiđamerkurjökull, Iceland. He postulated a positive feedback between bed irregularities that caused streaming of debris and locally concentrated bed erosion in grooves that, in turn, enhanced the bed irregularities causing further streaming of debris. Grooves are not common on granodiorites in the area but can be found in many distal areas on metamorphic rocks of the Shoo Fly Formation. This may represent the nature of basal debris carried over the two types of rock. Granodiorite tends to erode by plucking of large jointed blocks that are comminuted into rounded clasts, while metamorphic rocks of the Shoo Fly break are entrained as small, hard angular fragments. Conversely, grooves in the Shoo Fly rocks may represent their lower position in the glacial system, where meltwater and high hydrostatic pressures play a greater role.

Fig. 4. Nailhead striae on slate bench eroded into north side of Monumental Ridge in Lake Valley. Ice flowed from bottom to top.

3.2. Crescentic gouges

Small arcuate fractures on brittle glaciated rock surfaces, collectively referred to as *friction cracks*, have long been known in the geomorphic literature (Chamberlin, 1888). They are typically small, ranging from a few centimeters up to 50 cm (Fairbridge, 1968). Two types of friction cracks are most common: *crescentic fractures* with horns pointing down-ice and *crescentic gouges* with horns pointing up-ice. In both cases, a low-angle plane dips (relative to the rock surface) in the direction of ice flow (Gilbert, 1906a; Harris, 1943; Benn and Evans, 1998). Crescentic gouges often have a steep arcuate face dipping about 70° up-ice and a second fracture surface dipping 20° down-ice (Harris, 1943).

Crescentic gouges are by far the most common type of arcuate rock fracture in the area. Numerous large crescentic gouges near Big Bend in the South Yuba Canyon extend from the valley bottom to ~ 100 m up the valley side. They range in size up to 2 m (linear horn-to-horn) in the valley bottom where ice was more than 300 m thick (Fig. 5). Although small crescentic gouges are commonly described as occurring in longitudinal rows, these are rarely preserved on granodiorite in the study area. Large crescentic gouges may cluster in a particular locale and occasionally occur side-to-side as pairs in double crescents, but not in a longitudinal array. Gilbert (1906a) described similar extremely large crescentic gouges on granodiorite rocks in the central Sierra Nevada. He explained that they formed by two pressure fields: an oblique pressure established a conoid fracture of percussion and a direct pressure set up a conchoidal fracture. A rock entrained in the base of the glacier exerts a great forward and downward pressure focused at a point in the bedrock. The underlying rock is elastically deformed in front of the boulder and bulges until the elastic limit of the rock is exceeded. A brittle fracture occurs and a sliver of rock is dislodged from the intersection of the two fracture surfaces (Gilbert, 1906a). Harris (1943) performed laboratory experiments and found that the size of crescentic gouges increased with the surface area of the contact. The largest gouges in the study area occur in the deepest position of the valley, however, suggesting that ice depth may somehow influence gouge size.

3.3. P-forms

A suite of subglacially sculpted bedrock surface features known collectively as *P-forms* have been defined based on various shapes and orientations including sichelwannen, flutes, and potholes (Benn and Evans, 1998). The nature of subglacial meltwater erosion and its importance to glacial erosion has been a subject of considerable debate, and P-forms are often evoked as evidence of subglacial meltwater erosion.

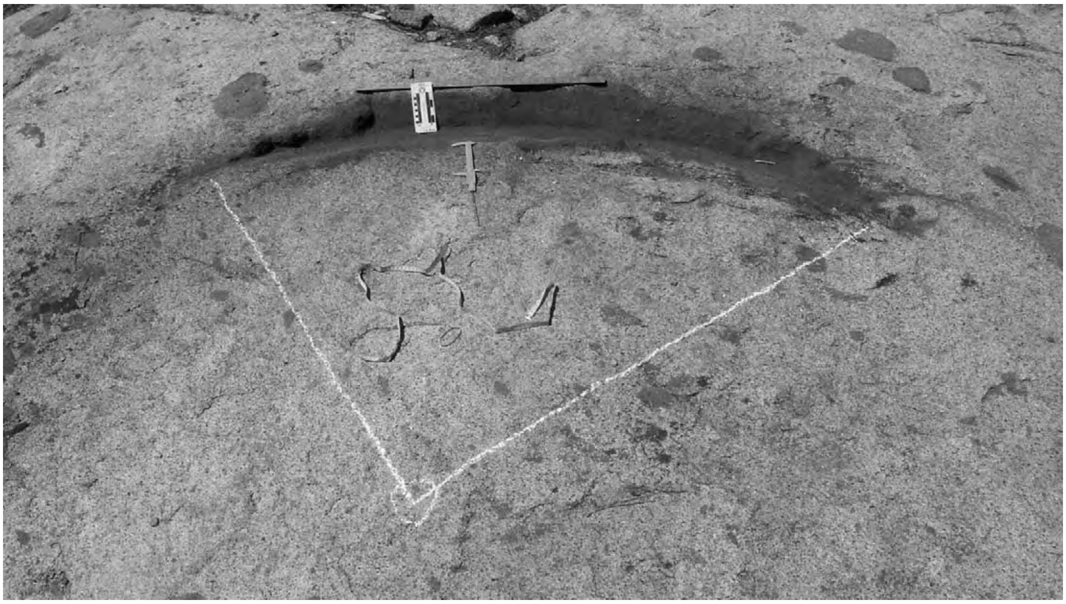

Fig. 5. Large crescentic gouge in granodiorite at bottom of South Yuba Canyon. Radius ~ 134 cm, maximum depth ~ 9.6 cm. Vertical scale bars in centimeters (left) and inches (right).

Some writers have likened P-forms to fluvially eroded bedrock forms and argue that subglacial erosion alone is responsible for them (Shaw, 1994). Some of these features have been observed in association with till in subglacial environments, however, and combinations of fluvial and glacial processes may be involved in the formation of some P-form features. Boulton (1974) argued that sichelwannen result from streaming of debris concentrations as was discussed under the formation of grooves. In deep subglacial environments where P-forms tend to be located, saturated till under high pressure may behave like a plastic.

Temperate glaciers often have elaborate subglacial drainage systems with hydraulic pressures that vary substantially at a variety of time scales (reviewed by Willis, 1995). For example, subglacial hydrostatic pressures may increase in the early spring melt season due to closure of ice tunnels over the preceding winter (Hooke et al., 1985). Glacial sliding velocities and erosion rates may be strongly related to subglacial hydrostatic pressures, so subglacial drainage conditions can exert considerable influence on local subglacial processes that may be responsible for P-forms.

Between a pair of bedrock benches in the lower Bear Valley where ice flow was constricted, a series of shallow, smoothly scalloped, undulating surfaces are developed on fine-grained metasedimentary rocks of the Shoo Fly Formation. These nondirectional features may be incipient potholes but are so shallow that horizontal forces were clearly responsible for their sculpture. Similar features have been described in the central Sierra (Gilbert, 1906b, Fig. Y) and explained as the result of migrating moulins that trained meltwater from crevasses onto rocks beneath the glacier. Crevasses may have been located in this extensional environment where ice spilled through the constriction, but moulins are ephemeral features in such an environment and not necessary for the generation of subglacial meltwater. Gilbert's example included large potholes in close association with the shallow scalloped surfaces; but no potholes occur at the Bear Valley site and attempts to explain the two types of features are best separated unless a link can be demonstrated. These forms are not striated in spite of their basal position in a valley constriction and the lack of weathering as evidenced by fresh polish. This suggests that either the basal layer was starved of coarse sediment or hydrostatic pressures were sufficient to lift the basal layer above its bed. Abundant erratics in the area suggest that coarse sediment was not in short supple.

Given the low topographic position within a major valley constriction, the role of subglacial meltwater or high hydrostatic pressures are suspected to have played an important role in forming the streamlined undulating features in the Bear Valley constriction, either by direct meltwater erosion or by rapid basal sliding. Numerous P-forms are also located at the north end of Bear Valley and in the entrance to Yuba Gorge. Substantial hydraulic heads are likely to have developed at both of these locations that were the lowest points and served as outlets for subglacial meltwater from the Spaulding ice field.

3.4. Riegel, rock bars, and knock-and-lochain topography

Riegel are bedrock ledges oriented transverse to the prevailing ice-flow direction. Conventionally, the term has been used to describe well-spaced, individual ridges at the lower rims of rock basins or cirque floors. They are commonly associated with the breaks in stair-stepped valleys shown in standard geomorphology references (Matthes, 1930, p. 90; Fairbridge, 1968, p. 467; Flint, 1971, p. 128). Cotton (1942) argued that individual riegel can be caused by truncation of valley spurs, structurally induced differential glacial erosion, or vertical glacial corrasion. While present in the NW Sierra, riegel are not the dominant transverse erosional features in the area.

Extensive clusters of closely spaced transverse ribs across the granodiorite floors of the Fordyce, Spaulding, and Castle ice fields range in height from 5 to 20 m (Fig. 6). The individual ridges of rock will be referred to here as rock bars for lack of a better term. They are more closely spaced and lack the asymmetry of riegel. They are structurally controlled and oriented parallel to master joints approximately perpendicular to the ice-flow direction. These rock bars are relevant to glaciology and glacial geomorphology in at least two regards. First, they represent a very high bed roughness that resisted basal sliding. Any attempt to model basal sliding, shear stress, or erosion in this area

Fig. 6. Granodiorite rock bars in floor of Fordyce Canyon at south base of Old Man Mountain where valley glacier entered upper Spaulding ice field. View to NNW across north-striking master joints (from lower left to upper right) exerting strong structural control on knock-and-lochain topography. Ice flow from lower right to upper left.

must incorporate a high roughness for these surfaces. Second, these rock bars appear to reflect the manner in which the jointed granodiorite rocks in this area erode, that is, by plucking of large blocks between the remaining rock knobs. Analysis of the spacing of rock bars and joints may reveal a periodicity or pattern representing a stable bed form under the most erosive conditions of deep ice. Boulton (1974, p. 68) described theoretical interrelationships between bedform wavelengths, amplitudes, and ice dynamics. A thorough analysis of the mechanics of these processes is beyond the scope of this paper but could elucidate fundamental processes and rates of glacial erosion over much of the Sierra granitic terrain.

Large areas of rough bedrock terrain have been referred to as *knock-and-lochain topography* from Gaelic words for knoll and small lake (Linton, 1963; Benn and Evans, 1998). While this term is not common in the North American glacial geomorphology literature, it describes the large areas of eroded granodiorite beneath the Fordyce, Castle, and Spaulding ice fields where the local relief of large areas is dominated by rock bars.

3.5. Roches moutonnées

Roches moutonnées are asymmetric forms with streamlined stoss sides and steep, rugged lee sides. They are common throughout the Sierra Nevada at a variety of scales ranging from a few meters to more than 100 m tall. Roches moutonnées have been described in the literature as ranging up to 150 m high, while larger asymmetrical hills up to 350 m high are sometimes referred to as *flyggbergs* (Benn and Evans, 1998). The asymmetry that characterizes these landforms is caused by abrasion on the gentle stoss slopes and plucking on the lee sides. Jahns (1943) concluded that roche-moutonnée formation is dominated by lee-side erosion that is much more rapid than abrasion on the stoss side.

Plucking on the lee side of large roches moutonnées requires rock fracturing that may occur by any of three mechanisms: frost shattering, wedging of rock fragments, and subglacial water pressure variations (Sugden et al., 1992). Water pressure fluctuations not only induce rock fracture but also encourage entrainment of fractured rock material. High water pressure in lee-side cavities is associated with reduced overburden pressure, freezing onto rock fragments, increased shear stress, and reduced frictional resistance. Rapid reductions in cavity water pressures increase stress gradients in the bed and encourage joint propagation (Iverson, 1991b). Meltwater can be delivered to lee-side cavities through transverse crevasses that form in the extensional environments common in these positions (Hooke, 1991), or large hydraulic heads may develop through elaborate subglacial conduits.

Asymmetry is enhanced by pressure differentials in the longitudinal direction. Asymmetry tends to be best expressed under thin ice where the ice overburden pressure is low so the longitudinal pressure differential is maximized (Benn and Evans, 1998). Thin ice also facilitates the introduction of meltwater to lee side cavities. In theory, therefore, roches moutonnées and similar asymmetrical features are best expressed under thin ice. Under deep ice, symmetrical forms such as whale backs are more likely to develop. The environment in which large roches moutonnées form has been the subject of much debate (Sugden et al., 1992). Some have argued that erosion is greatest during maximum glacial periods (Sugden and John, 1976), while others have argued for the greatest erosion rates during periods of growth or ablation (Boulton and Clark, 1990).

A pair of large roches moutonnées eroded in mafic crystalline rock impeded flow of the valley glacier at the bottom of the upper South Yuba Canyon (Fig. 2). They are strongly asymmetric, with stoss-side ramps of Hill 6642 and Cisco Butte extending 500 and 400 m, respectively, and lee-side faces extending only 200 and 240 m, respectively (Fig. 7). These hills rise to between 120 and 150 m above the bench on which they rest, ~ 300 m above the floor of South Yuba Canyon. Lee faces are extremely steep, jagged, and over-deepened below the elevations of the plateau on the stoss sides. The tops of both hills have abundant fresh striae. Striae on Hill 6642 extend right up to the edge of the lee-side cliff face attesting to the effectiveness of abrasion on the hill top and a lack of flow separation until the abrupt face. The flat hill crest extends about 50 m with very little lowering near the sheer lee-side face.

Two cosmogenic radionuclide surface-exposure ages averaging 13.4 ± 740 obtained from an erratic on Hill 6642 (James et al., 2002) indicate that Tioga ice overtopped this hill. Mapping of the Tioga maximum

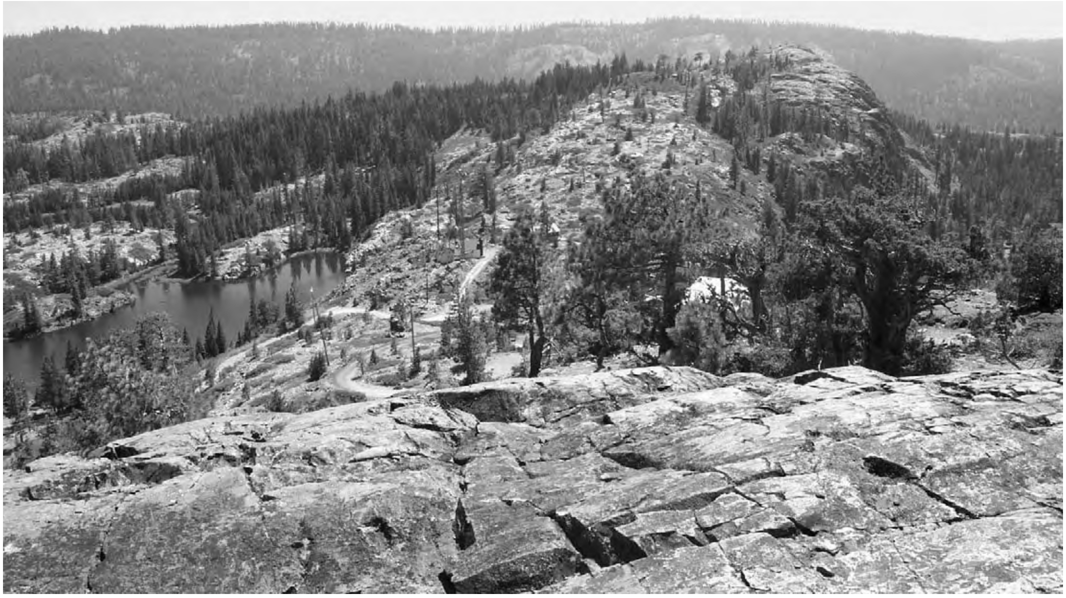

Fig. 7. Hill 6642 roche moutonnée at mouth of upper South Yuba Canyon. View to south from top of Cisco Butte, another roche moutonnée with an abraided stoss side (foreground). Ice flow from left to right.

elevation on nearby valley sides indicates that the maximum height of the Tioga glacier was not deep over the top of Hill 6642 or Cisco Butte, probably less than 30 m thick above the hill tops. At least two other glaciations as high but not much higher than the Tioga ice occurred in this valley, so these large roches moutonnées were apparently created under thin ice during multiple glacial maxima. The location of these roches moutonnées on a high surface above the main valley and away from valley sides suggests that meltwater was not introduced by deep subglacial conduits. Thin ice over the hill-tops may have resulted in transverse crevasses and probably facilitated meltwater introduction to the lee sides from above.

Old Man Mountain is a flybberg or very large roche moutonnée with a steep lee side on an asymmetrical peak rising more than 500 m above the floor of Fordyce Canyon. Tioga glaciers did not overtop Old Man Mountain (James et al., 2002), but swept around both sides of the peak at a high elevation and may have supplied meltwater to the lee side. One or more pre-Tioga glaciers probably overtopped Old Man Mountain.

A much smaller roche moutonnée east of Tuttle Lake indicates, along with striae and erratics that ice flowed across Rattlesnake Ridge north of Rattlesnake Peak into Rattlesnake Canyon. Numerous smaller roches moutonnées are scattered throughout the region and form a continuum of landforms of varying asymmetry and girth grading to whalebacks and rock benches.

3.6. Whalebacks

Whalebacks are streamlined rock knobs with symmetrical longitudinal profiles caused by abrasion of both their stoss and lee sides. In theory, whalebacks tend to form in areas of deep ice with rapid flow velocities. Small whalebacks can form under only a few hundred meters of ice, but larger examples form under deep ice streams (Evans, 1996). Abrasion on the lee side of a whaleback indicates a lack of separation of the ice from the bed at least some of the time. Thick ice or a low-viscosity basal layer tend to suppress bed separation (Evans, 1996). If flow separation occurs, over-steepening or plucking on the lee side may be suppressed by lack of water pressure variations within the lee-side cavity. This may result from cold, thick ice preventing meltwater from reaching the bed.

Whalebacks in the study area are found on valley bottoms where valley glaciers were quite deep. For example, a large granodiorite whaleback on the floor of the upper South Yuba Canyon at the Big Bend ranger station is in the deepest part of the canyon. The whaleback is about 25 m high, 80 m long, and 50 m wide with large crescentic gouges on both the stoss and lee sides. It presumably formed under thick ice and was not altered to an asymmetrical form by shallow ice during the period of ablation and ice thinning. Lack of erosion by thin ice does not preclude recessional ice in these locations, but it suggests that no prolonged shallow glacial flows were effective in altering these features.

3.7. Lateral benches and ice-marginal channels

High on many valley walls in the NW Sierra, eroded bedrock benches are graded longitudinally in a manner very similar to lateral moraines. They were initially enigmatic because erosional benches have not been previously described in the Sierra Nevada. In several locations, these benches are the dominant valley-side morphological features and they can easily be mistaken for lateral moraines from a distance. They have little or no till cover and are high on valley walls, sometimes more than 300 m above the valley floor (Fig. 8) and are often graded at or near the paleo-ice surface. They may extend a few hundred meters but rarely more than that. Bench widths range from more than 30 m wide to zero where the benches pinch out against valley walls or end abruptly. Bench tops are relatively flat in the cross-valley dimension, even when developed on steep hill slopes, and their proximate slopes meet valley walls at an over-steepened cut slope.

Several eroded benches are located on S-facing exposures, but only a few have been found on N-facing slopes. They often occur on the up-ice terminations of ridges where ice bifurcated around the ridge. These positions were presumably subject to high basal shearing, as well as contributions of water and sediment from upslope. A boulder line is sometimes found 5–15 m above the bench indicating ice this thick deposited erratics above the bench. The ubiquity, clear relation to glaciation, and lack of previous recognition of valley-side benches in the Sierra calls for a review of possible explanations for these erosional features.

Cotton (1942, pp. 286–299) devoted a chapter to multiple-benched valley-side profiles and reviewed possible explanations including structural controls, small troughs eroded into large troughs, lateral moraine terraces, and epiglacial benches (a.k.a., ice-mar-

Fig. 8. Erosional bench high on steep flank of Rattlesnake Mountain (off photo to right). Interpreted as erosional remnant of ice-marginal channel. Flat striated surface eroded into vertically dipping layered metamorphic rocks of Sailor Canyon Fm. Floor of South Yuba Cn. is straight ahead more than 300 m down.

ginal channels). Structural controls can be ruled out because the Sierra benches develop gentle gradients parallel to ice surface slopes in many kinds of rock, including vertically dipping layered metamorphics (Fig. 8), granodiorites with varying joint orientations, and massive volcaniclastics. Explanations that involve valley deepening into a previously broader valley (cf. Garwood, 1910) do not apply here. The benches are flat surfaces with fresh lateral incision into valley walls and are too young to represent an old, high, valley-side surface. Erosion by basal sliding along the glacial margin has been well documented. On the Austerdalsbreen, Glen and Lewis (1961) measured ice margin slippage of 26 cm/day or 65% of the velocity of the maximum centerline velocity. Yet, this process does not explain the flat bench floors, nor has such a process-form link been suggested in the literature.

Only one plausible mechanism in the glacial geomorphology literature can account for the features observed in the study area; that is, ice-marginal or submarginal channels. Although little stratified drift has been found in association with these surfaces that may be due to a lack of exposures, subsequent erosion, or burial by supraglacial till or colluvium. Tarr (1914) described ice-marginal and submarginal channels, sketched a cross section of a flat bedrock bench eroded on a steep valley side by such a process, and cautioned that they should only be used for mapping as minimum ice-surface elevations due to the possibility that they formed in submarginal positions. Ice-marginal and submarginal channels up to 2 km in length and graded along valley sides have been described by Price (1973). Ice-marginal channels occur at the subaerial contact between ice and bedrock. Both the ice surface and the valley side slope toward the contact so water that does not seep into the glacier tends to flow toward the channel. If the downvalley gradient of the contact is steep, the channel will tend to cut into the ice and leave no evidence at the ice margin (Price, 1973). If the channel gradient is gentle, a bench can be cut into the hill slope, especially where the slope is mantled by weak rock. Flint (1971) argued that marginal channels will only form on slopes of moderate gradient, and Maag (1969) suggested that steep valley-side slopes may be associated with the formation of benches rather than channels. Sugden and John (1976) described ice-marginal channels that eroded simple flat benches in hillsides.

The eroded lateral benches in the study area are interpreted as ice-contact channels. They indicate minimum ice elevations because they may represent submarginal channels with the ice surface above. This ice surface can often be located by boulder lines above the bench and in many cases this suggests that the channels were not deep below the ice surface. While basal sliding fails to explain the flat top surface, basal erosion of a pre-existing marginal or submarginal channel could explain the apparent lack of stratified drift. A polygenetic explanation may be appropriate given that several glacial advances occurred in these valleys. The apparent preferred S-facing aspect of these features may reflect seasonal differences in surface snow melt and supraglacial runoff. Runoff earlier in the melt season would not penetrate the ice surface as easily as later in the season when subglacial channels are enlarged and better connected. Thus, early thaw snow-melt runoff from south-facing slopes may have been associated with larger ice-marginal flows than later runoff from north-facing slopes that percolated down into the glacier.

3.8. Crag and tail

Crag and tail is a Scottish expression for a resistant knob that obstructs ice flow with a tapered tail in the protected lee zone (Fairbridge, 1968). Most commonly, these are small features with tails of erodible or deformable basal till, but the tail also may be streamlined weak bedrock. Small erosional crag-and-tail features were described by Chamberlin (1888, p. 193) as flow-direction indicators. The landform is scale independent, however; and Castle Rock, Edinburgh, with the streamlined Royal Mile, has often been described as a large crag and tail (Fairbridge, 1968; Benn and Evans, 1998; Martini et al., 2001). The Edinburgh crag is a volcanic plug with a bedrock tail extending about 1.4 km (Fig. 9). Bedrock mapping below a thin but variable till sheet indicates that the Edinburgh plug stands about 110 m above a horseshoe-shaped frontal trough in the bedrock on the up-ice side (Sissons, 1971; Evans and Hansom, 1996). This suggests severe scour and high ice velocities in front of the crag.

Devil's Peak is a large crag and tail on the divide separating the South Yuba and North Fork American Canyons (Fig. 10). The crag protruded above the ice

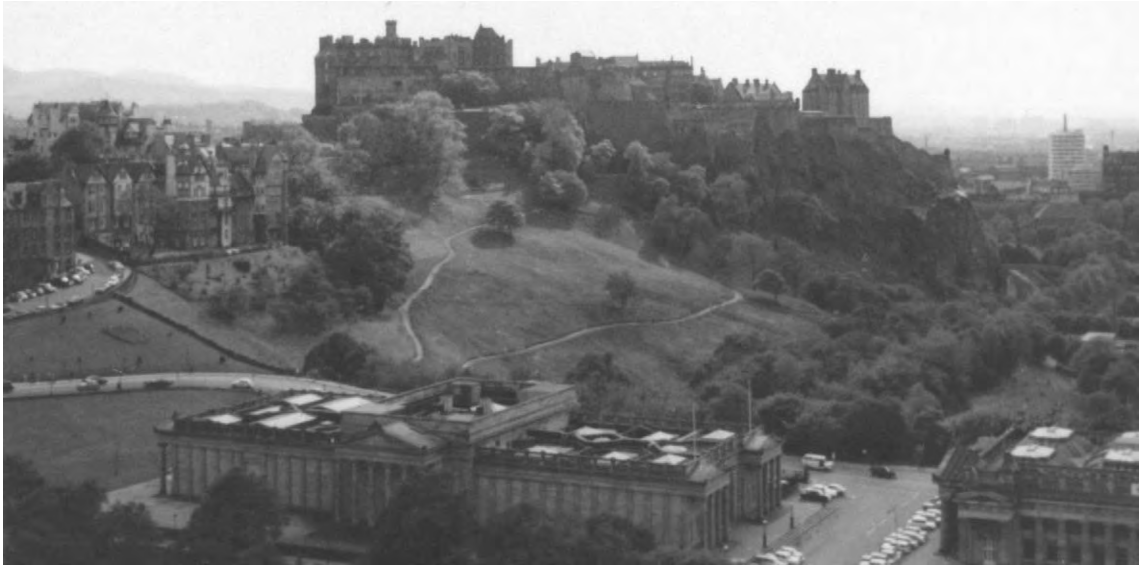

Fig. 9. Edinburgh Castle, Scotland, on crag and tail. Royal Mile extends along tail to left. Ice flow was upper right to lower left. Photograph by Alex Shepherd, 1973 (used with permission).

surface and is a Tertiary basalt capping andesite (Lindgren, 1897; Harwood, 1980). Harwood described Devils Peak as 170 m thick and composed of two basalt flows with strong columnar jointing. Hudson (1951) described nearby basalt plugs and flows in the Castle Peak area and concluded from thin sections that

Fig. 10. Devils Peak crag and tail. View ESE across upper South Yuba Cn. from Rattlesnake Ridge. Main ice flow was down South Yuba Cn. (middle left to lower right), but high, thin ice spilled across plateau (left to right around Devils Peak) into North Fork American Cn. beyond center horizon. Central Pacific Railroad on valley wall; Interstate 80 out of view in bottom of South Yuba Cn.

many of the extensive andesites mapped by Lindgren (1897) were Pliocene basalts lying disconformably on surfaces of low relief. Dalrymple (1964) obtained a K/Ar age of 7.4 My from a basalt on nearby Boreal Ridge.

The layered basalt in Devils Peak was resistant to glacial erosion while the tail, composed of weaker andesitic material, was streamlined. The front of Devil's Peak has no topographic trough. Streamlining of the tail was not complete and an asymmetric spur protrudes on the SW side (Fig. 11). Yet, this asymmetrical form should not be confused with horned crags described by Jansson and Kleman (1999), that are more symmetrical and formed under deep ice. While the asymmetry and lack of a trough suggest early stages of crag-and-tail development, the glacial streamlining of Devil's Peak represents substantial erosion. Glacial erosion is often assumed to be concentrated in valley bottoms but negligible on uplands (Clayton, 1965; Price, 1973). Yet, Devils Peak is on an upland plateau where ice was much thinner than in the main South Yuba Canyon. Effective erosion by shallow ice around Devils Peak was presumably enhanced by steep flow gradients into the deep Royal Gorge of the North Fork American Canyon where local relief is 1340 m at Snow Mountain. It also reflects the relative weakness of the andesite volcanics surrounding the crag and of the rhyolites forming the low plateau on the north side.

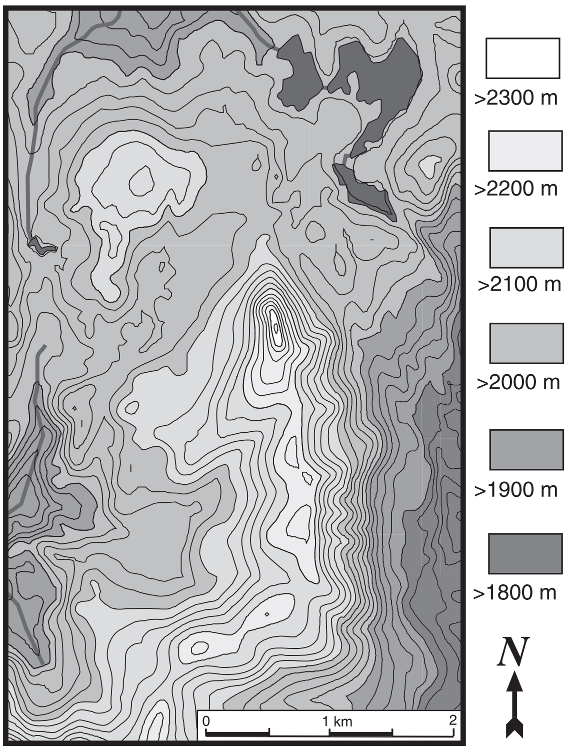

Fig. 11. Topographic map of Devils Peak (center) showing crag with streamlined tail to south. Ice flow from South Yuba Cn. (north of map), across plateau, into deep Royal Gorge (south of map).

3.9. Valley morphology

A reoccurring question in glacial geomorphology is the extent to which Pleistocene glaciers created the deep troughs that they occupied. Whether the classic U-shaped valley represents minor valley-bottom alterations or major valley deepening has been debated. Many examples of glaciated V-shaped valleys have been documented-particularly in lower reaches of glaciated valleys-that indicate limited glacial erosion of the fluvial form (Embleton and King, 1968). The South Yuba Gorge is an example of a V-shaped glaciated valley (James, 1996). On the other hand, examples of the ability of glacial ice to erode deep troughs are well known. For example, the bottoms of deep fiords below sea level and Paleozoic glacial troughs cut into areas of continental shield (Fairbridge, 1968, p. 459).

Evans (1997) advanced eight propositions regarding the effectiveness of temperate alpine glacial erosion. Two of the propositions are directly concerned with the effectiveness of glacial valley erosion: (i) glacial and related processes dominate the geomorphology of glaciated mountains and (ii) troughs are glacial paleochannels calibrated to the discharge of ice and the erodibility of bedrock. Harbor (1992) linked a two-dimensional, finite-element model with an erosion model and simulated the production of U-shaped valleys from V-shaped valleys. If the model was run for a long period of time, simulating repeated occupations of a valley by ice during the Quaternary, the U-shape was propagated downward and deep narrow troughs resulted.

Deepening of Sierra valleys by Pleistocene erosion was postulated by Small and Anderson (1995) based on an analysis of cosmogenic radionuclides in upland rocks that showed relatively little late Quaternary erosion. They suggested that deep glacial erosion concentrated in valley bottoms together with isostatic rebound have substantially increased local relief. Geomorphic field evidence qualitatively supports the concept of glacial valley deepening and relative stability of surfaces above the glacial limit.

The importance of glacial erosion to the present morphology of valleys in the region is difficult to prove, but evidence of severe glacial erosion is abundant. This evidence includes gouges, roches moutonnées, whalebacks, and other features described above, the U-shaped cross-section morphologies of valleys such as Bear Valley (cf. James, 1996), and evidence of erosion from analysis of cosmogenic radionuclides (Fabel et al., in review). Where large valley glaciers flowed out of ice fields, they left characteristic trough forms. Exceptions can be found where valleys do not display typical glacial form, and this may involve combinations of limited ice flow and structural weaknesses in the rock.

Rattlesnake Canyon provides an excellent example of sudden change in glaciated valley morphology across a bedrock contact. The middle reaches of Rattlesnake Canyon are developed in granodiorite and are characterized by wide cross sections with numerous rock bars. Glacial ice widened the base of the granodiorite portion of the valley and subsequent fluvial erosion has not substantially lowered the valley floor. In contrast, the lower valley is developed in slate and is steep walled with a deep V-shaped bottom. Rattlesnake Creek has cut a series of waterfalls in the slate beginning immediately downstream of the granodiorite contact. The cataracts of lower Rattlesnake Canyon are the result of fluvial incision that has modified the shape of the valley cross section by incising a V-shaped notch in the valley bottom. Rapid fluvial incision may have been encouraged by a lowered base level at the mouth of the canyon. South Yuba Canyon at the mouth of Rattlesnake Canyon appears to have been deepened by late Pleistocene glaciers into a narrow trough more than 300 m deep.

4. Sedimentary products of glacial erosion

The widespread glacial erosion in this area clearly generated a tremendous volume of sediment, yet the landscape at high elevations is dominated by bare rock surfaces. While till deposits and occasional lateral moraines remain in the area, the overwhelming dominance of eroded bedrock attests to the efficiency of sediment removal from the area and its transport downvalley. Outwash terraces are poorly preserved in the mountainous gorges of the NW Sierra, although occasional remnants can be found, as in the South Yuba Gorge near the town of Washington and in the North Fork American River at Green Valley and between Ponderosa and Long Point. Most glaciofluvial sediment has been eroded from the mountain canyons to low-gradient reaches of the rivers in the Sacramento Valley and beyond. As with the glacial record, the alluvial record from the Sierra is better understood to the south in the San Joaquin Valley than to the north in the Sacramento Valley.

Alluviation in the Central Valley has largely been in response to glaciation in the Sierra Nevada (Marchand and Allwardt, 1981). For example, the upper member of the Modesto Formation in the San Joaquin Valley has at least four terraces interpreted as *Tioga* outwash deposits (Marchand and Allwardt, 1981). Central Valley alluvial cycles are often assumed to have been synchronous not only with Sierra Nevada glaciations, but also with global continental ice (Marchand and Allwardt, 1981; Dupré et al., 1991). Strict synchroneity between alpine glacial events and sea level changes that integrated global changes in continental ice should not be assumed, however (Gillespie and Molnar, 1995). Alpine glaciers respond relatively rapidly to minor climate fluctuations while continental glaciers have a large thermal inertia that requires an extended period of climate change to overcome. Subtle differences in timing between alpine glaciers and sea-level changes may have strongly influenced alluvial sequences downvalley in basins influenced by coastal base levels.

During Sierra glacial advances aggradation by outwash left deposits now in alluvial terraces grading out from foothill fan areas. In distal zones, however, lower sea levels during global full-glacial periods caused degradation. During Sierra glacial recessions and interglacial periods, sediment production was reduced and outwash near the mountain front was incised. Lower in the system, however, rising sea levels during global interglacial periods encouraged aggradation. These complex spatial and temporal interactions point to the need for higher resolution glacial and fluvial chronologies to develop process–response linkages between climate change and glacial and fluvial landforms.

Atwater and Belknap (1980) attributed thorough erosion of Pleistocene transgressive estuarine deposits to erosion during low eustatic cycles, implying sediment production was substantially reduced before sea-level rise was complete. Much of the sediment from the severe glacial erosion documented in this paper was probably delivered downstream rapidly while sea levels were still relatively low. Retreat of late *Tioga* ice from Fordyce Canyon was complete by ~ 14,100 ^{26}Al YBP (James et al., 2002), about the time that global sea levels began accelerating around 14,000 YBP (Bard et al., 1996). Thus, relatively efficient flushing of glacial sediment may have been well underway before marine transgression was completed. This may have reduced alluvial deposition in floodplains and estuaries and facilitated degradation of alluvium in the Sacramento Valley.

5. Conclusion

The NW Sierra Nevada provide a diverse set of landscapes varying from high alpine granodiorite terrain of moderate local relief to extremely rugged, deeply incised gorges of the North Fork American River. Great contrasts exist between the inaccessible but widely admired Wild and Scenic North Fork American River Canyon area, the heavily traveled but geomorphically ignored Interstate 80 corridor down the South Yuba Canyon, and the relatively unstudied isolated roadless area to the north. Much can be learned from glacial geomorphology, and these landscapes exemplify processes dominated by the sculpture of glacial ice. Evidence of glacial valley deepening and relative stability of surfaces above the glacial limit is abundant. In contrast, some upland areas covered by relatively thin ice show indications of severe erosion.

Some of the glacial landforms described in this report have not been previously identified in the Sierra Nevada. The interpretation of high, eroded, lateral benches as ice-marginal channels should be treated as a hypothesis for thorough field testing. They are potentially important indicators of former minimum ice elevations that are otherwise difficult to map because of poor preservation of moraines. Similarly, implications of rapid and erosive thin ice, drawn from the Devils Peak crag and tail, are preliminary and should be tested with thorough field mapping to determine the number of glaciations and thickness of ice in that area and to the south. While extensive lateral moraines support these interpretations, only reconnaissance mapping has been done there.

The fundamental relationships between glacial landforms and inferred glacial processes bear few surprises. Whalebacks and crescentic gouges are largest and best expressed in deep ice positions. Roches moutonnées of various scales appear to have formed in positions under thin ice. Large former ice fields are now characterized by rugged granodiorite knock-and-lochaine topography with numerous structurally controlled rock bars. Zones where ice discharge was the greatest are characterized by deep troughs. These generally display the classic U-shape with the exception of a few locations where dominance by fluvial or glaciofluvial incision is apparent, such as in lower Rattlesnake Canyon and the South Yuba gorge.

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