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SOIL GENESIS AND MORPHOLOGY OF A MONTANE MEADOW IN THE NORTHERN SIERRA NEVADA RANGE

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Given the importance of riparian areas in the western United States, knowledge about the spatial distribution, properties, and genesis of these soils is surprisingly limited. In conjunction with an interdisciplinary study of the impacts of grazing on soils and vegetation, we characterized three pedons along a hydrologic gradient on a montane meadow of the northern Sierra Nevada range. Radiocarbon dating of charcoal indicates that meadow pedogenesis began approximately 3600 years B.P., after a catastrophic valley erosional event. Since that time, nearly 1 meter of soil has accumulated over a basal glaciolacustrine unit. Critical factors and processes influencing soil genesis and morphology include: seasonal variation in soil redox status, frigid soil temperatures, additions of volcanic tephra, wildfires, and polygenesis related to Holocene climatic, hydrologic, and vegetation changes. Argillans are present on ped faces of certain soil horizons, which suggests extended dry periods at which time clay pervection occurred. Clay mineralogy is disjunct; surface horizons are dominated by kaolinite and underlying horizons by smectite. The high clay content of such youthful soils suggests rapid primary mineral weathering. Charcoal-containing strata attest to frequent wildfires during the Holocene epoch. The spatial complexity of soil patterns and their properties infers that these riparian areas are dynamic, and their character may have been shaped by previous climatic patterns.

In xeric climates, riparian ecosystems play a critically important role in water quality, stream flows, wildlife habitat, and forage production. Previous work in more mesic areas has documented the ability of riparian buffer strips to absorb nutrients from upland runoff (Lowrance and Leonard 1988; Haycock and Pinay 1992). Thus, properly functioning riparian zones help improve the water quality of streams. Riparian zones may also store water during wet periods that can be released to the associated stream during dry periods, thereby improving minimum flows (Elmore and Beschta 1987). Wildlife tend to use riparian areas disproportionately more than other habitats (Thomas et al. 1979), and these areas are also an important source of forage (Reid and Pickford 1946; Roath and Krueger 1982).

Assessment of management impacts and short-term trends requires knowledge of the forces that influenced the formation and dynamics of riparian ecosystems. Unfortunately, few studies have characterized soil properties, the spatial variability of soils, and pedogenesis in these systems. A multidisciplinary team was assembled to evaluate the response of vegetation, soils, and water quality to grazing in a montane meadow of the northern Sierra Nevada range, California. The purpose of this paper is to report on the properties and pedogenesis of soils on a montane meadow of the northern Sierra Nevada range. The record of these soils documents a varied pedogenic history and underscores the difficulty in management of these complex ecosystems.

MATERIALS AND METHODS

Site characteristics

The study area is along the upper reach of Grizzly Creek, a tributary to Lake Davis and the Feather River (Figs. 1 and 2). At the research site, the floodplain is approximately 100 m wide, but width varies considerably. The surrounding uplands are forested by lodgepole pine [*Pinus contorta* ssp. *murrayana* (Grev. & Balf.) Critchf.], white fir [*Abies concolor* (Gordon & Glend.) Lindley], and Jeffrey pine [*Pinus jeffreyi* (Grev. & Balf.)]. The dominant vegetation on the meadow consists of Kentucky bluegrass [*Poa pratensis* (L.)], baltic rush [*Juncus balticus* (Willd.)], Nebraska sedge [*Carex nebrascensis*

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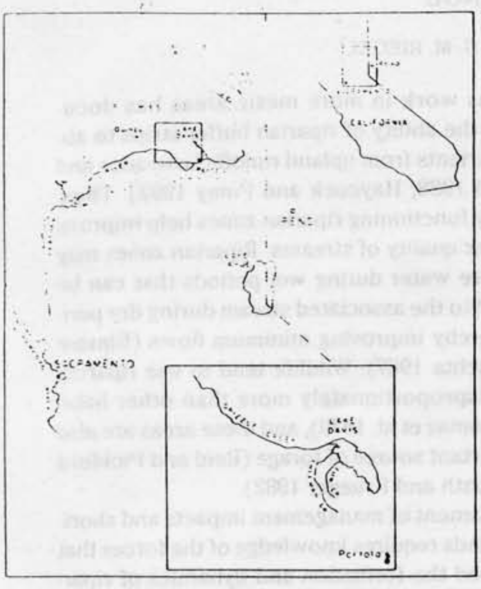


FIG. 1. Location map of study area.

(Dewey)], and tufted-hairgrass [*Deschampsia caespitosa* (L.) Beauv.]. The species composition differs systematically from stream edge to forest boundary: Nebraska sedge dominates very poorly drained areas along the stream, other graminoids increase in poorly drained midfloodplain positions, and forbs increase in the better drained floodplain-upland interface.

Elevation of the site is approximately 1800 meters. Precipitation at nearby Three Mile Valley averaged 89 cm per year (SD = 34) from 1959 through 1991, approximately 85% of which is in the form of snow falling in the months of November through March. The geology of Grizzly Creek extending through Lake Davis is mapped as Quaternary lacustrine deposits (Burnett and Jennings 1962). The geology of the surrounding area is complex. Adjacent uplands are mapped as Miocene, Oligocene, and Pliocene pyroclastics. Minor associated rock-types include Miocene andesite and Mesozoic granodiorite. Mapped along the upper reaches of Grizzly Creek is Paleozoic marine limestone and dolomite.

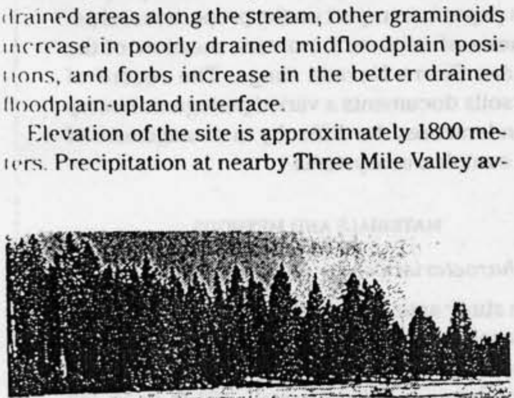


FIG. 2. Photograph of Grizzly Creek meadow taken October 22, 1994. Usually, cool fall temperatures decrease plant water demands sufficiently, thereby raising water levels in the stream. This fall, however, the stream remains dry, vividly demonstrating the extent of a 7-year drought.

In 1990, soil oxidation-reduction potentials were measured weekly from June 5 through August 16, biweekly from August 29 through October 2, and monthly through November 15. In 1991, redox was measured weekly from May 27 through August 13, and biweekly through October 9. Platinum wire (1.5 cm) soldered onto steel welding rods was used. Pairs of redox probes were placed at 20- and 40-cm depths at sampling points located at 2.5, 10, 20, and 50 m on a transect line from the stream to the forest boundary. A PVC salt bridge was located between the two redox probes to facilitate measurements under dry conditions. Readings were taken using a portable millivolt reader with an AgCl reference electrode. Water table depths were measured at the same locations. In 1990, water table was measured weekly from May 4 through August 16. In 1991, water table was measured twice a month from February 29 through April 22, weekly from May 15 through August 18, and biweekly from September 5 through October 9. Perforated 5-cm PVC observation wells were placed to a depth of 1.5 m. Soil thermistors were placed at depths of 10, 60, and 120 cm in the midfloodplain soil. Daily maximum, minimum, and average temperatures were recorded via a datalogger from August 1993 through July 1994.

A hydraulic coring device was used to examine the soil around the research area to a depth of approximately 3 m. Based on this information, three sites within the research plots were sampled for detailed soil description and characterization: stream edge just beyond a subtle levee (10 m from stream), midfloodplain (30 m from stream), and the floodplain adjacent to the forest boundary (50 m from stream). To minimize site disturbance, small approximately 0.5-m tesseras were excavated. Pedons were described and sampled

TABLE 1

Pedon descriptions

Horizon	Depth (cm)	Texture	Matrix color moist	Matrix color dry	Primary structure	Argillans
Stream edge						
Oa	0-10	sicl	10YR 2/1	10YR 4/2	weak, medium, granular	absent
A	10-28	sicl	2.5Y 2/0	10YR 4/1.5	moderate, fine & medium, granular	absent
AB	28-38	sic	2.5Y 2/0	10YR 5/1	weak, medium & fine, prismatic	absent
Bw	38-61	sic	5Y 2.5/1	10YR 5/1	strong, medium, prismatic	few ped faces; few pores
2Cg	61-79	sil	2.5Y 6/2	10YR 8/1	weak, coarse, platy	absent
3Btgb	79-99	sic	5Y 3/1 & 5/1	10YR 5/1 & 7/1	weak, coarse, prismatic	few ped faces; common pores
3Cg	99-112+	sicl	5Y 5/2	5Y 7/1	massive[a]	absent
Midfloodplain						
Oa	0-10	sil	10YR 2/1	10YR 4/2	moderate, medium, granular	absent
A	10-20	c	10YR 2/1	10YR 4/1	weak, coarse, subangular blocky	absent
AB	20-28	c	2.5Y 2/0	10YR 4/1	moderate, fine & medium prismatic	absent
Bw	28-46	sic	2.5Y 2/0	10YR 4.5/1	strong, fine & medium, prismatic	few ped faces; few pores
2C	46-58	sicl	10YR 4.5/2	10YR 7.5/1	weak, medium & coarse, prismatic	absent
3Btgb	58-81	sic	2.5Y 2/0	10YR 6/2	strong, medium, prismatic	few ped faces; common pores
3Cg	81-107+	sic	5Y 5/2	5Y 6.5/2	massive[a]	absent
Forest edge						
A	0-10	scl	10YR 2.5/1	10YR 4/1	moderate, fine & very fine, granular	absent
AB	10-23	scl	10YR 3/2	10YR 5/1.5	strong, fine & very fine, granular	absent
Bt1	23-41	scl	10YR 3/1.5	10YR 5/2	moderate, coarse, prismatic	many ped faces
Bt2	41-66	l	10YR 4.5/2	10YR 6/2	moderate, coarse, prismatic	few ped faces
2Btb	66-97	cl	5Y 3.5/1	2.5Y 6/2	moderate, medium, prismatic	many ped faces
2Cg	97+	sic	5Y 5/2	5Y 6/2	massive[a]	absent

[a]Has a varved depositional structure with occasional coarse strata.

using established procedures (Soil Survey Staff 1984).

Laboratory

Soils reserved for laboratory characterization were immediately air-dried in the field and then transported to the laboratory. Soil was crushed lightly to pass a 2-mm sieve and stored in paper cartons. All subsequent analyses were done on this material. Soil pH was measured in 0.01 M $CaCl_2$ (McLean 1982). Organic carbon was measured by the Walkley-Black procedure (Nelson and Sommers 1982). Total N was quantified by the Kjeldahl method (Bremner and Mulvaney 1982). Particle size determination used established procedures of Gee and Bauder (1986). Sand, silt, and clay fractions were saved for later analyses. Cation exchange capacity was measured by the 1 N NaOAc at pH 7.0 method (Soil Survey Staff 1984). Exchangeable cations were quantified with an atomic absorption spectrophotometer. Moisture desorption properties were quantified using a pressure plate apparatus (Richards and Fireman 1943). Quantification of wet aggregate stability used the method of Harris (1971). Various pools of Fe and Al were extracted using citrate-dithionite, acid ammonium oxalate, and pyrophosphate (McKeague and Day 1966; Soil Survey Staff 1984).

The <2-μm size fraction, reserved after particle size separations, had the following treatments performed before X-ray diffraction (Moore and Reynolds 1989): K-saturation and air-drying; K-saturation with heating to 150°C for 15 min; K-saturation with heating to 450°C for 1 h; Mg-saturation followed by vapor intercalation with glycerol; Mg-saturation followed by vapor intercalation with ethylene glycol. Clays were sedimented on glass slides before X-ray diffraction. X-ray diffractograms were run using the following conditions: Phillips automated diffractometer system PW1710, Bragg Brentano goniometer equipped with curved graphite diffracted-beam monochromator, 0.2-mm receiving slit, incident and diffracted beam soller slits, divergence and antiscatter slits, Cu K alpha radiation, take-off angle 6°, step size of 0.02 degrees with step counting time of 0.5 s, at 0.04 degrees/s, theta compensating slit. The coarse-silt through medium sand fractions were examined with a petrographic microscope and minerals identified using standard methods (Stoiber and Morse 1981; Phillips and Griffen 1981). Particle percentages were determined on the very fine sand fraction using the line count method (Brewer 1976). One hundred grains were counted for each soil horizon.

In 1991 a large pit was excavated near the east edge of the study area to obtain charcoal for radiocarbon dating. The position of the pit was slightly lower and wetter than the other studied pedons and on the side of an older stream meander approximately 5 m north of present Grizzly Creek and approximately 100 m downstream from the stream edge pedon. Normally, the water table would have been close to the soil surface at this position, but 1991 was a very dry year after several below average precipitation years. We chose strata for ^{14}C dating based on the presence of sufficient charcoal. The strata from which the charcoal was extracted showed no evidence of disturbance. Charcoal was sent to a commercial lab (Beta Analytical Inc., Miami, FL) for radiocarbon dating.

RESULTS AND DISCUSSION

Site information

In the 4 years of this study, there was remarkable variation in precipitation. On average, snow is completely off the meadow by late March to late May. Subsequent to snow melt, the entire width of the meadow is flooded. Redox potentials during this period are low, ranging between −231 and 627 mV at 20-cm depth and −249 to 598 mV at 40-cm depth. Even during this soil saturation period, redox potentials closer to the forest are more positive as a result of their slightly higher position. Plant growth proceeds rapidly after snow melt, and except for the low bench astride the stream, surface soils become significantly more aerated. By early fall, redox potentials ranged from 89 to 724 mV near the stream to 301 to 722 mV near the forest edge.

In early May of 1990, after snow melt and initial water table draw down, water table depth was 31 cm at the stream edge, 45 cm at mid-floodplain, and 50 cm at the forest edge. Water table draw down was very gradual at the stream edge, dropping to 46 cm on June 28. A dramatic decline began on July 18, with depth dropping to 97 cm and reaching 114 cm by July 31. The mid-floodplain and forest edge sites dropped further by July 5, to 105 and 108 cm, respectively. By July 31, both sites were approaching 135-cm depth. A similar pattern was determined for 1991. Caution must be used in extrapolating from these data because 1990 and 1991 were drought years.

Soil thermistor readings from August 1993 through July 1994 indicate a frigid soil temperature regime. The stream has a gradient of 1 to 2% and would be classified as a (C with an E potential) based on Rosgen (1994).

Pedon descriptions

The stream edge and midfloodplain pedons have similar horizonation (Table 1). The surface O horizons have a weak to moderate granular structure that, when wet, is very unstable and smears easily. The A horizons are fine-textured and have granular to blocky structure that readily smears when wet. The presence of clay cutans (argillans) in the field and distinct prismatic structure suggests a Bt horizon. Upon microscopic examination, however, the suspected clay cutans were very faint. Moreover, given that these soils are seasonally frozen, compression and compaction during freezing may force thin films of water containing suspended fine clay to orientate on surfaces of peds (Gorbunov 1961). Thus, we classify these horizons as Bw. The Bw horizons overlie a light-colored, platy-structured horizon that, at the time the pedons were excavated, had lateral water movement occurring toward the stream. Based on this information, the layer was field identified as an E horizon; however, the very fine sand fraction was dominated by volcanic tephra with plentiful diatom frustules (Table 2, Fig 3, A, B, and C). It is clear this layer is a lithologic discontinuity and is thus identified as a 2C horizon. Below the tephra layer is a fine-textured, gleyed horizon with pronounced prismatic structure and distinct, thick, clay cutans. The cutans were deposited via percolating water as suggested by a waxy luster, by the fact they can be peeled off when dry, and by flow structures on the cutans. The stratigraphic position of this horizon and its appearance suggests it is a paleosol B

TABLE 2

Line counts (in percent) of the very fine sand fraction for the collected pedons

Horizon	Glass	Glass-like[a]	Quartz & Feldspar	Amphiboles Pyroxenes[b]	Phytoliths	Freeze agglomerates[c]	Others[d]
Stream edge							
Oa	37	8	28	9	3	10	5
A	17	32	12	3	23	10	2
AB	19	22	15	6	7	22	10
Bw	41	9	21	3	3	22	0
2Cg	92	2	1	2	0	1	1
3Btgb	15	10	50	3	1	13	10
3Cg	9	0	67	19	0	5	1
Midfloodplain							
A	9	0	55	11	4	19	3
AB	15	0	33	11	3	32	6
Bw	23	0	24	18	0	29	6
2C	76	4	3	4	2	11	0
3Btgb	18	8	33	4	0	32	4
3Cg	15	10	50	3	1	13	10
Forest edge							
A	13	3	48	11	2	18	7
AB	14	1	40	13	0	30	1
Bt1	12	0	41	10	0	37	0
Bt2	25	3	14	8	0	47	3
2Btb	2	0	29	5	0	62	2
2Cg	not sampled						

[a] A semitransparent material that appears as devitirified glass but often has enmeshed diatom tests.
[b] Green hornblende is dominant.
[c] Particles that visually appear as weathered biotite but are actually Fe-rich halloysite that agglomerate together to form large optically uniform domains (Blank et al 1992)
[d] Includes zircons, opaques, and diatoms.

horizon, which we identify as 3Btgb. Underlying this horizon is a gleyed and strongly mottled horizon. The horizon has a varved depositional texture with occasional coarse-strata, especially nearer the stream. We interpret this layer to be the Quaternary-aged glaciolacustrine unit mapped by the U.S. Geological Survey (Burnett and Jennings 1962).

The slightly higher forest edge pedon is distinctly different from the midfloodplain and stream edge pedons. An O horizon is absent, possibly because of the relative dryness of this location. The Bt horizon has much more pronounced structure and more plentiful clay skins than the Bw horizons of the midfloodplain and floodplain soils. The 2C horizon is absent in the forest edge pedon; however, the transitional zone between the Bt2 and 2Btb horizons shows streaks of whitish material. Moreover, the Bt2 horizon is especially rich in volcanic glass (Table 2). The paleosol 2Btb horizon has more pronounced and plentiful clay films than occur in the 2Btgb horizons of the stream edge and midfloodplain pedons. The underlying varved lacustrine sediments occur as well in the forest edge pedon (Table 1).

Chemical and physical properties

Selected chemical and physical attributes are presented in Table 3. Soil pH is 5.0 or less in surface horizons and increases with depth. The pedons are all very high in organic carbon in surface soil horizons, but this gradually decreases with depth. Carbon to N ratios are less than 20 for soil above the basal lacustrine U. Clay content is remarkably high, far higher than that estimated in the field; organic matter apparently masks the high clay content. Sand content is highest in the forest edge pedon. High organic matter and silt and clay content foster elevated moisture retention capacity, especially in the O and A horizons of the stream edge pedon. Cation exchange capacities are very high and generally decrease with depth. Given the high organic matter content of surface horizons, a portion of the cation exchange capacity is attributable to pH dependent charge; the pH of the extracting solution (7.0) is greater than the natural soil pH. The exchange complex of all pedons is dominated by calcium, followed by magnesium, then sodium and potassium. Percent base saturation is less than 50% in surface horizons and increases to greater than 60% in the paleosol Bt horizons and below. Index of soil aggregation, which has ramifications for streambank stability, is generally lowest in the stream edge pedon.

Pools of extractable iron are very low (Table 4). Anoxic conditions promote the formation of the soluble Fe^{+2} species which is then mobilized from the system. Dithionite-extractable Fe and Al are highest in surface horizons and generally decrease with depth (Table 4). The pyrophosphate extraction, which measures organic bound Fe and Al, is, as expected, highest in the organic-rich horizons. Oxalate-extractable Fe generally decreases with depth, then increases in the paleosol Bt horizon; oxalate-extractable Al generally decreases with depth (Table 4). The ratio of dithionite-extractable Fe to Al correlates with the presence of the paleosol Bt horizons and underlying lacustrine sediments (Table 4).

Soil stratigraphy-age relationships

The stratigraphy of the charcoal dating pit closely matches that of the characterized pedons. The A horizon is thicker in the charcoal dating pit, possibly because of greater sedimentation. The tephra rich 2C horizons of the stream edge and midfloodplain pedons were not found in the charcoal dating pit; however, a stratigraphically equivalent stratum contained lapilli with some volcanic ash.

From the radiocarbon dates obtained, soil genesis on the meadow began after 3600 ± 90 yrs B.P (Table 5). The paleosol Bt horizon is at least older than 2840 ± 220 yrs B.P. and not older than 3600 ± 90 yrs B.P. Soil material above the paleosol Bt horizon, including the volcanic tephra, is younger than 2840 ± 220 yrs B.P.

Mineralogy

The importance of volcanic tephra in the genesis of these soils is evident from line counts of the very fine sand fraction (Table 2; Fig. 3 A and B). Tephra amounts were greatest in the 2C horizons of the stream edge and midfloodplain positions. The purity of tephra in these horizons and presence of admixed diatom frustules (Fig. 3C) suggests a quiet water environment at the time of deposition. Hydraulic soil coring of the meadow indicates the tephra layer extends more than one-quarter km above and below the study area. We attempted to determine the provenance of the glass by matching its chemistry to dated tephra deposits; unfortunately, the glass segregated into four chemical groups. It appears that the magma that produced the tephra was either differenti-

TABLE 3

Selected chemical and physical properties of the pedons

Horizon	pH	% Organic C	C:N	Particle size distribution			Moisture retained (MPa)			CEC[a]	Exchangeable				% Base saturation	IA[b]
				sand	silt	clay	0.005	0.10	1.5		Ca	Mg	K	Na		
				g kg^{-1}						cmol kg^{-1}						
Stream edge																
Oa	4.82	11.4	19.7	147	536	317	963	512	318	57.3	13.7	2.7	0.77	1.20	32.1	50
A	5.35	8.9	10.6	88	558	355	1523	694	368	78.5	12.5	2.7	0.33	1.20	21.3	52
AB	5.52	4.0	11.7	47	463	489	825	553	315	54.6	16.8	4.6	0.72	1.41	43.1	38
Bw	5.66	2.5	16.1	34	508	458	648	429	259	44.0	17.5	4.3	0.64	1.38	54.2	29
2Cg	6.00	0.7	18.9	169	669	162	720	262	110	14.1	3.8	0.8	0.53	0.77	41.9	16
3Btgb	5.73	1.0	15.6	73	475	452	668	377	241	46.5	19.4	7.4	0.74	1.08	61.4	20
3Cg	5.89	0.6	27.7	197	516	287	588	291	155	28.2	11.4	4.3	0.63	0.78	60.8	90
Midfloodplain																
Oa	5.00	15.5	19.1	117	637	246	876	503	380	59.8	13.4	3.8	0.99	0.98	32.1	21
A	5.29	5.6	12.7	162	289	549	551	360	259	55.2	16.6	5.0	0.46	0.97	41.6	43
AB	5.39	3.0	9.6	144	308	547	526	328	241	43.5	17.2	5.4	0.41	1.01	55.3	65
Bw	5.58	1.8	13.9	105	449	446	548	327	216	41.5	15.8	5.2	0.42	1.15	54.4	79
2C	5.96	0.9	11.2	168	554	279	781	311	175	28.4	9.9	2.8	0.30	1.05	49.6	31
3Btgb	5.69	1.1	8.9	91	520	389	568	325	208	42.5	19.3	7.4	0.53	0.95	66.2	75
3Cg	5.83	1.0	31.1	105	557	338	687	351	186	25.2	14.6	5.2	0.52	1.27	85.8	78
Forest edge																
A	4.83	5.0	16.2	510	217	273	474	240	174	30.2	8.4	2.4	0.31	0.97	40.0	48
AB	4.92	2.0	11.7	529	197	274	340	168	134	23.4	8.8	2.4	0.24	0.93	52.9	26
Bt1	4.96	1.7	12.4	507	219	274	343	185	126	19.7	8.2	2.3	0.22	1.13	60.2	23
Bt2	5.25	0.7	15.8	436	362	203	462	213	116	18.2	11.0	2.8	0.22	1.00	82.6	19
2Btb	5.87	0.7	11.3	286	413	301	542	255	148	28.7	14.7	5.9	0.32	1.01	76.5	55
2Cg	not sampled															

[a]CEC, exchangeable bases, and % base saturation all determined by the pH 7.0 ammonium acetate method (Soil Survey Staff 1984).

[b]IA or index of aggregation is a measure of wet aggregate stability (Harris 1971). Values near 100 imply extreme stability; values near zero imply poor stability.

ated and zoned or that quiet water conditions persisted for a time such that several different tephras were incorporated (Bruce Cochran 1990, personal communication).

A glass-like material was plentiful in the very fine sand fraction, especially in the stream edge pedon (Table 2; Fig. 3 D and E). Particles are an admixture of semiopaque, yellowish-brown, isotropic material with striations of weakly birefringent material. A refractive index of approximately 1.46 suggests the isotropic material is opaline silica. By focusing the microscope in and out, minute diatom frustules are seen encased in the siliceous material. We believe this material represents silica deposited around roots in response to plant water uptake, a mechanism which has been reported previously (Buurman et al. 1973). Diatoms, which are a component in the wet environment, die and become encased in the silica precipitate.

Other abundant particles in the very fine sand fraction include "freeze agglomerates" (Fig. 3F). The appearance of these particles suggests a weathered biotite; however, several factors mitigate against this conclusion. First, its large proportion in some samples is incompatible with the low levels of biotite in surrounding rock types. In addition, X-ray diffraction indicates that halloysite is the principal mineral comprising freeze-agglomerates (Blank et al. 1992). We hypothe-

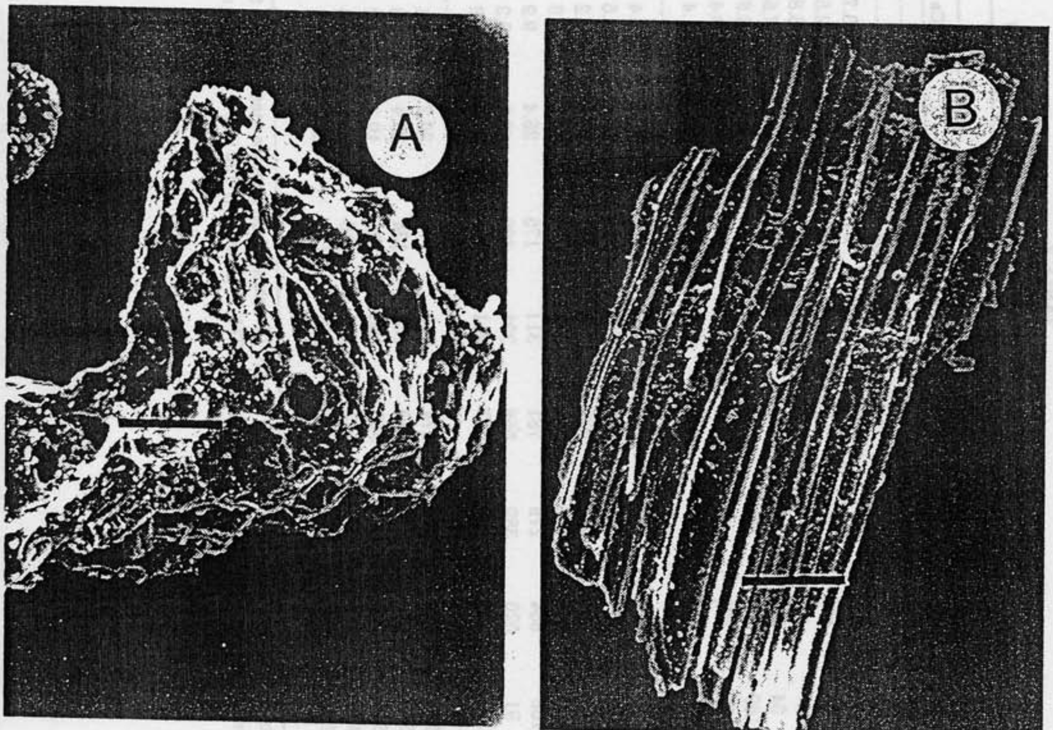
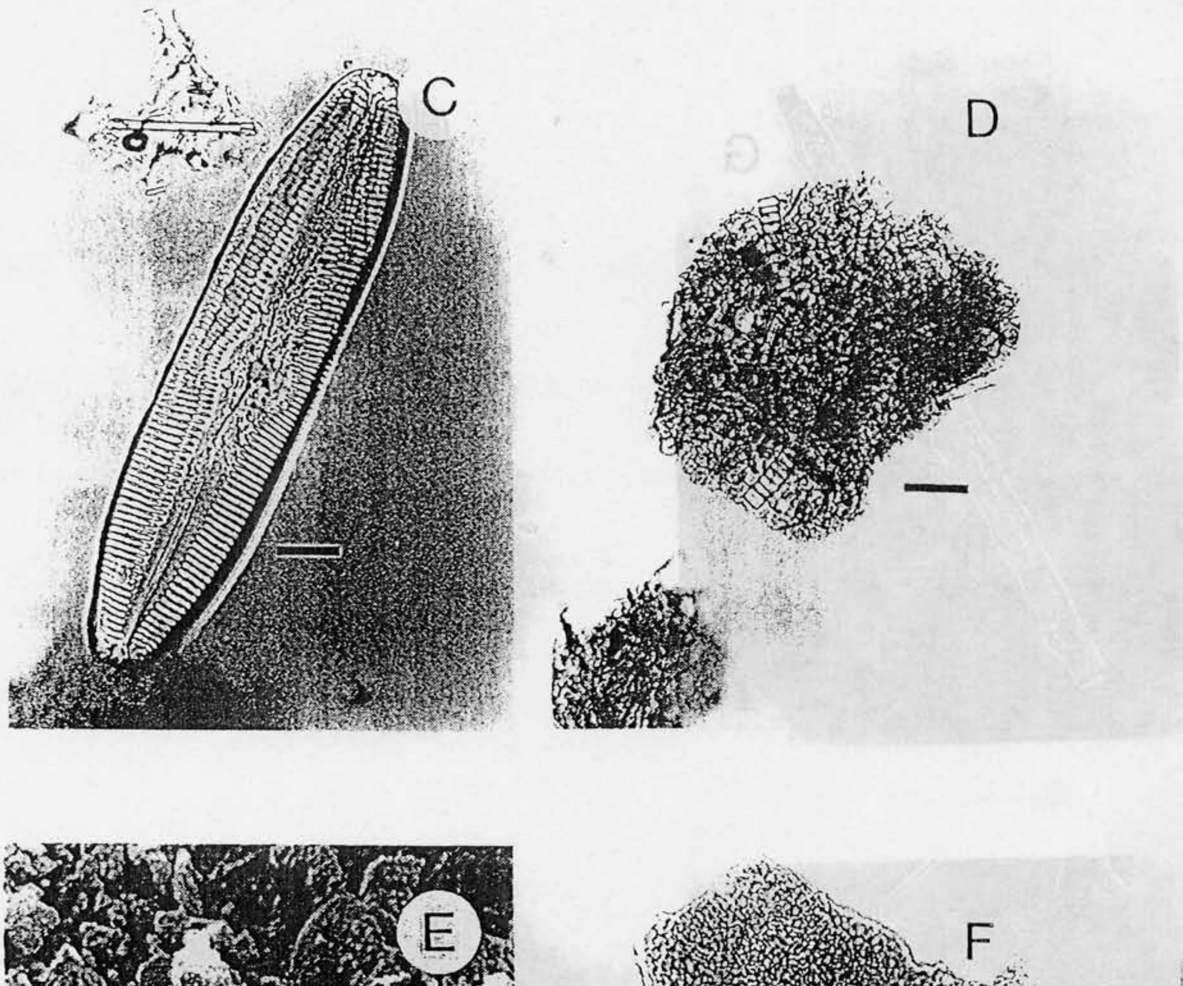


FIG. 3. Photomicrographs of selected soil particles. Scanning electron micrographs of the two most common pumiceous shards extracted from the 2Cg horizon of the stream edge pedon. One type has spherical to oval vesicles (A) and the other elongated tubular vesicles (B). Photomicrograph (C) is a frustule of the most prevalent species of diatom admixed with the pumiceous shards. A "glass-like" particle is very common in the A horizon of the stream edge pedon. Particles are isotropic, translucent, yellowish-brown in color, and often have minute diatom frustules encased in the isotropic material (D). Viewed with the scanning electron microscope, the "glass-like" particle appears made of up smaller particles pasted together (E). Freeze-agglomerates make up a sizable portion of the mineral fraction of the meadow soils. This sample, from the 2Btb horizon of the forest edge pedon, shows the typical microgranular surface texture and plate-like form (F). A myriad of phytolith forms occur in most soil horizons. In the A horizons of the stream edge pedons, the abundance of phytoliths is striking. Forms range from rod-like forms (G) to joined cell-like bodies (H). The majority of phytolith forms occurring in the present A horizon are far different from that of *Carex nebrascensis* (I) or *Juncus balticus* (J). All line scales = 20 μm, except for E which is 10 μm.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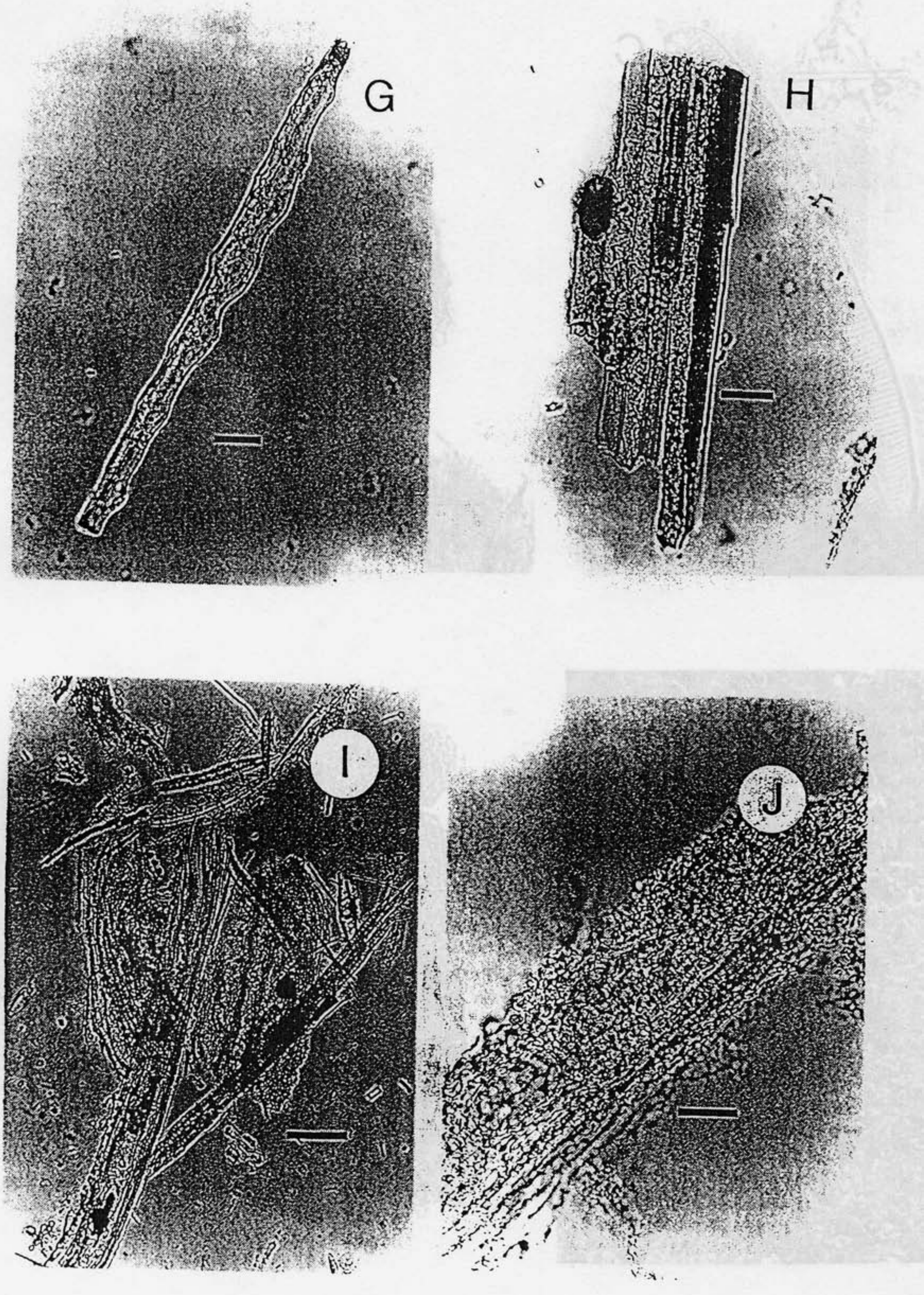
G
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J

sized that these particles are formed via freeze-induced accretion of Fe-rich colloidal halloysite (Blank and Fosberg 1991). The formation of freeze-agglomerates may increase soil physical stability.

In certain soil horizons, particularly the A horizon of the stream edge pedon, phytoliths are plentiful. The most common morphologies evident were as long cylindrical units (Fig. 3G) and as quadrilaterals composed of cell-like units (Fig. 3H). Phytolith (G) is possibly a template of xylem or other cylindrical-shaped cells, and (H) is likely from epidermal cells (Mulholland and Rapp Jr. 1992). We also extracted plant silica from the leaves of *Carex nebrascensis* and *Juncus balticus*, which now dominate the vegetation of the stream edge pedon. Plant silica from *Carex nebrascensis* is dominantly in the form of quadrilaterals in which individual fibers are evident (Fig. 3I), whereas phytoliths of *Juncus balticus* consist of more massive epidermal secretions that are not composed of subunits (Fig. 3J). The morphological dissimilarity between plant silica of present meadow species and soil phytoliths from the A horizon suggests (i) the stream edge position was occupied by different vegetation in the past and (ii) overland flow or stream flooding may deposit phytoliths from areas occupied by different vegetation.

The clay mineralogy of the studied pedons is disjunct (Fig. 4). Soil horizons above the paleosol 2Btb and 3Btbg horizons are dominated by kaolinite; the 2Btb horizons and horizons below contain appreciable smectite. The youngest soil horizons have prominent reflections at 0.72 nm and 0.36 nm that correspond to the 001 and 002 planes of kaolinite (Bailey 1980). The lack of a strong reflection at 0.445 nm (02 and 11 planes) indicates that halloysite is not a major component. In the Sierra Nevada range west of Reno, NV, halloysite is the dominant clay mineral under pine forests (Birkeland 1969; Birkeland and Janda 1971). Potassium-saturated and air-dried treatments show a weak reflection at 1.02 nm, which expands near 1.4 nm with magnesium and glycerol intercalation; thus, vermiculite is a minor component. Given the youthfulness of these soils, the high clay content and dominance of kaolinite in the clay fraction is surprising. Plant activity can greatly increase mineral weathering kinetics (Mortland et al. 1956; Robert and Berthelin 1986). Perhaps the meadow vegetation fosters rapid mineral weathering and clay formation. A variation of ferrolysis is another possibility (Brinkman 1970). Ferrolysis occurs when seasonal reduction causes Fe^{+2} to replace other cations on exchange sites. As the water table lowers later in the season, the Fe^{+2} is oxidized, and clay surfaces are acidified. Perhaps this seasonal acidification promotes hydrolysis of primary minerals. The high content of rapidly weatherable volcanic glass in these soils may, in part, explain the high clay content. However, given the youthfulness of these soils, it is unlikely that tephra would have weathered significantly to clay minerals, and if the tephra had weathered appreciably, there should have been far higher levels of oxalate-extractable Al attributable to the formation of short-range ordered minerals (Lowe 1986). However, one cannot discount a possible interaction between volcanic tephra and the meadow environment that fostered rapid weathering without the production of short-ranged ordered minerals.

The paleosol horizons and underlying glaciolacustrine sediments also contain appreciable kaolinite (Fig. 4). Potassium-saturated and air-dried samples have a reflection at 1.0 nm that expands to between 1.5 and 1.68 nm with Mg and glycerol intercalation. Magnesium and ethylene glycol intercalation (not shown) increases this spacing to between 1.73 and 1.74 nm and allows a small 1.4-nm reflection to be more easily deciphered. Smectite is, therefore, a major component, and vermiculite a minor component, of the clay-sized fraction. The increase in the 001 spacing with glycol over that of glycerol suggests the smectite may be beidellite (Borchardt 1989). Birkeland and Janda (1971) did not identify montmorillonite within soil profiles of the Sierra Nevada of roughly equivalent temperature and precipitation of the Grizzly Creek study site.

Smectite and kaolinite are co-stable (Kittrick 1969). High solution levels of silicic acid favor smectite stability, whereas lower levels favor kaolinite. In surface soil horizons, spring snowmelt may flush silicic acid, thereby favoring kaolinite stability. Underlying horizons, which in most years remain wet or near saturation, are not leached. Soil-solution levels of silicic acid may remain high enough to foster either neoformation of smectite or prevent inherent smectites from weathering to kaolinite. High levels of volcanic glass may also contribute to higher solution levels of silicic acid.

TABLE 4

Chemical extractions and ratios of the extractions of the collected pedons[a]

Horizon	Fe_d	Al_d	Fe_p	Al_p	Fe_o	Al_o	Fe_o/Fe_d	Fe_d/Al_d	Fe_p/Al_p	Fe_o/Al_o
	---------- g kg^{-1} ----------									
	---------- Stream edge ----------									
Oa	5.80	2.13	1.98	3.83	4.39	5.80	0.76	2.72	0.52	0.76
A	3.24	4.11	0.86	4.18	1.04	8.14	0.32	0.79	0.21	0.13
AB	2.82	0.89	0.61	3.07	0.81	4.55	0.29	3.17	0.20	0.18
Bw	2.29	0.73	0.37	1.73	0.57	3.22	0.25	3.14	0.21	0.18
2Cg	0.68	0.28	0.03	0.15	0.11	1.86	0.16	2.42	0.20	0.06
3Btgb	6.09	0.34	0.44	0.45	1.36	0.97	0.22	17.9	0.98	1.40
3Cg	3.59	0.26	0.09	0.22	0.50	0.65	0.14	13.8	0.41	0.77
	---------- Midfloodplain ----------									
Oa	4.57	1.03	1.35	2.42	2.53	3.15	0.55	4.44	0.56	0.80
A	4.49	0.92	1.17	2.74	1.81	2.13	0.40	4.88	0.43	0.85
AB	5.60	0.92	0.89	1.71	1.70	2.23	0.30	6.09	0.52	0.76
Bw	7.12	0.76	0.46	0.45	1.66	1.52	0.23	9.37	1.02	1.09
2C	2.65	0.48	0.08	0.25	0.50	2.30	0.19	5.52	0.32	0.22
3Btgb	5.11	0.53	0.14	0.16	1.22	1.40	0.24	9.64	0.88	0.87
3Cg	12.28	0.40	0.21	0.20	1.37	0.81	0.11	30.7	1.05	1.69
	---------- Forest edge ----------									
A	5.84	0.71	1.16	1.87	3.04	1.64	0.52	8.23	0.62	1.85
AB	5.67	0.52	0.51	0.42	1.70	1.24	0.30	10.90	1.21	1.37
Bt1	6.08	0.45	0.60	1.04	1.31	1.06	0.22	13.51	0.58	1.24
Bt2	5.51	0.35	0.11	0.15	1.00	0.90	0.18	15.74	0.73	1.11
2Btb	5.94	0.33	0.10	01.2	1.03	0.63	0.17	18.00	0.83	1.63
2Cg	not sampled									

[a] Subscripts d, p, and o refer to citrate-dithionite, pyrophosphate, and acid ammonium oxalate extractions, respectively.

Soil classification

The diagnostic surface epipedons for all pedons are umbric, based on dark colors, high organic C content, and base saturation less than 50%. Only the stream edge pedon has sufficient clay increase from the A to Bt horizon to qualify as argillic. Because nearby well drained upland soils developing on glacial till require nearly 40,000 years to form a textural B horizon, (Birkeland and Janda 1971), it is unlikely that the stream edge pedons meet the modal concept of an argillic horizon and more likely that they represent a finer depositional U. Similar reasoning makes it unlikely that paleosol Bt horizons qualify as argillic. In most years the soil moisture regime for all pedons would be aquic. The 2C horizons, in our opinion, do not fit the concept of an albic horizon. Based on these diagnostic properties, the pedons would be classified as Typic Humaquepts.

Reconstruction of soil genesis

The basal unit at the Grizzly Creek study site is a Quaternary-aged glaciolacustrine U (Burnett and Jennings 1962). There is no published information on the longevity of this lake or when the lake emptied. After lake recession, pedogenesis occurred on this basal U, if not during the Pleistocene, certainly in the drier early Holocene. Whatever soil development occurred during this time has been obliterated. Radiocarbon dates indicate that the soils are younger than 3600 years; the meadow associated with Grizzly Creek is very recent in origin.

Mountain meadows in the central Sierra Nevada underwent catastrophic valley cutting during the Holocene (Wood 1975). The impetus for this valley cutting is unclear; it could have been an intensely erosive precipitation event, a wildfire that destroyed protective vegetation, or a rare combination of factors. Ely et al. (1993) correlated large floods in the southwest between

480 and 3600 yrs B.P. with El Nino events. Once valley cutting begins, however, it leads to nearly complete removal of valley fill (Wood 1975). Toulomne Meadows (2700 m), in Yosemite National Park, has basal charcoal-bearing strata at 3930 yrs B.P. (Wood 1975) and is thus a near time analog of valley cutting to Grizzly Creek.

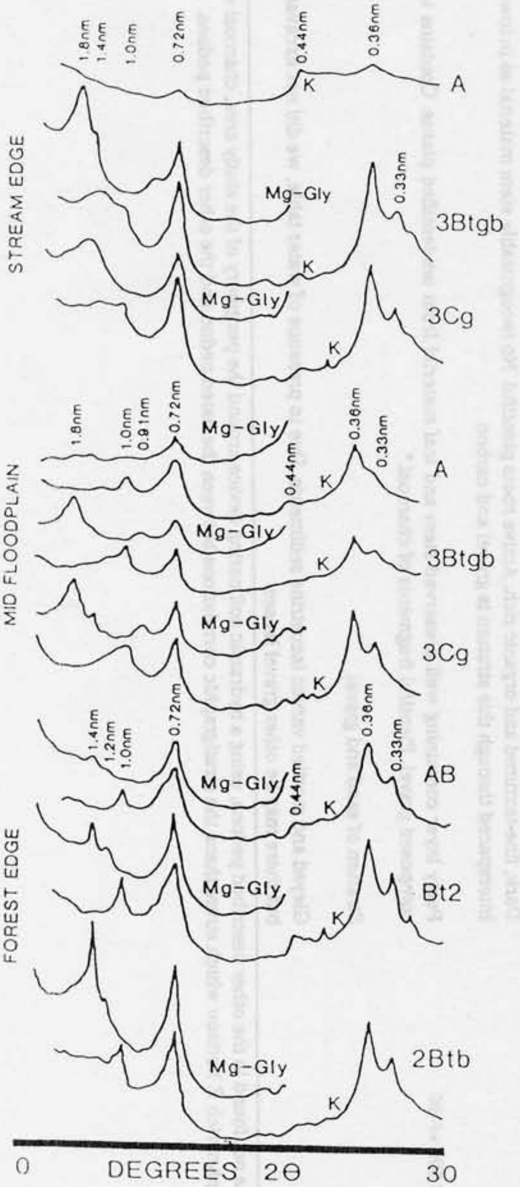

FIG. 4. X-ray diffractographs of clay-sized material from selected horizons. K-sat refers to potassium saturated treatment and air drying; Mg-Gly refers to magnesium saturation and vapor intercalation with glycerol. Only K-saturated treatment is included for the A horizon of the stream edge pedon.

After valley cutting, boggy meadow soils often occur at base level (Wood 1975), which is congruous with the peat layer at 160 to 170 cm in the charcoal dating pit. The depth of Holocene valley filling can be extraordinary, reaching several meters (Wood 1975). The texture of the valley fill deposits is dependent largely on the surrounding rock type and stream gradient. In the central Sierra Nevada, steep gradients and granitic rock types lead to coarse-textured soil deposits. In the Grizzly Creek riparian area, we conjecture that a gentle gradient, a preponderance of volcanic tephra, an interaction with meadow vegetation, and extrusive volcanic rock types lead to fine-textured valley fill sediments.

Pedogenesis of the initial valley fill sediment produced a differentiated soil profile in the time frame of 3600 yrs B.P. to until the time the lake-laid volcanic tephra was deposited. The paleosol Bt horizon formed during this interval. If hydromorphic conditions were prevalent at this soil-forming stage, little profile differentiation would have occurred because a high water table delimits pervection; the profile would have been an A-Cg type (Duchaufour 1977). The presence of clay skins on ped surfaces indicates that clay suspensions have percolated downward and oriented on ped faces (Buol and Hole 1961). Clearly the paleosol Bt horizon formed when the soil was freely drained. Many lines of evidence, including pollen cores, dendrochronology, and mountain meadow soil stratigraphy, suggest a dry period in the early Holocene—the altithermal (Antevs 1938; Wood 1973; LaMarche 1973; Anderson 1990). The exact dates of the altithermal are under debate, but it occurred between 9000 and 2800 yrs B.P. Wood (1975) determined that, during the altithermal, meadows in the central Sierra Nevada generally had good soil drainage and were forested; clay particles could pervect to form oriented coatings.

A major change in the hydrology of Grizzly Creek occurred sometime after 2840 yrs B.P. We believe this change was in response to pulses of cooler and wetter conditions in the Sierra Nevada — the neoglacial period (Curry 1969; LaMarche 1973; Burke and Birkeland 1983; Anderson 1990). The onset of neoglaciation raised the groundwater table in valleys, fostering the conversion of forests to meadows (Wood 1975).

In response to neoglacial cooling, the rising water table of Grizzly Creek formed a lake. It is

TABLE 5

Soil stratigraphy of radiocarbon pit

Horizon	Depth (cm)	Age B.P.	Comments
Oi	0–10		Vegetation dominated by *Carex nebrascensis* and *Juncus balticus*
A	10–40		Fine-textured, dark colored, with weak granular structure. No charcoal present. Measurements indicate root length of 1.5 m per cm^3 soil in the Oi and A horizons; this is extraordinarily high.
Bg	40–85		Fine-textured, prismatic-structured horizon which we conclude is stratigraphically equivalent to the most recent B horizons of the other collected pedons. Several large charcoal pieces were found, but sufficient quantity was not obtained for dating.
2Cg	85–110		Stratum of sand and gravel set in a fine and dark-colored matrix. Stratigraphy equivalent to the 2C horizons; however, thick tephra accumulations are not present at this location. Gravel sized material crushed easily with fingers and is dominated by glassy material.
3Btgb	110–130	2,840 +/–220	Strong prismatic structure with clay films on ped faces. Numerous fragments of charcoal throughout this layer. We interpret this layer to be stratigraphically equivalent to the paleosol Bt horizons of the other described pedons.
4Cg	130–137		Stratum of sand and gravel set in a fine and dark-colored matrix. Much of the gravel-sized material crushed readily between thumb and finger and has a whitish rind. X-ray diffraction of crushed gravel indicates the presence of albite and microcline but not quartz. It is possible the gravel-sized material is lapilli.
5Cg	137–160		Dark, fine-textured and organic rich. Active roots plentiful. No recognizable stem material as in lower peaty layer. Interspaced through this stratum is gravel and carbon.
6Oe	160–170	3,600 +/–90	Peaty layer containing well preserved stem and leaf material from unidentified plants. Contains intermixed decomposed gravel. Plentiful fragments of charcoal.[a]
7Cg	170–184		Stratum of sand and gravel.
8Cg	184+		Gleyed and mottled varved lacustrine sediments. Due to presence of water table, we did not excavate any further, but there may be other gravel lenses.

[a]Although carbon was not found in the other described pedons, using a hydraulic soil coring device around the periphery of the study area, charcoal was seen both above and below the tephra-rich E horizon which strengthens the stratigraphic correlations between the dated pedon and the other described pedons.

unknown if the rising water table alone was sufficient to create a lake or if factors such as log jams in the stream channel assisted. The lake served as a collector for volcanic tephra, thereby forming the 2C horizon above the old soil profile. The provenance and exact age of the tephra deposit eludes us at this time, but it must be younger than 2840 yrs B.P. Various eruptions from Mono and Inyo craters south of the study area fit this time frame, but it is doubtful ash falls from these eruptions extended north into our study area (Davis 1978; Sarna-Wojcicki et al. 1991).

There was another period of upward soil building following lake recession. Based on the present high water table, it is likely that soil thickening is still occurring. Pedogenesis on the most recent material has formed a B horizon with pronounced prismatic structure. Given the high water table of the meadow and that capillary movement of water maintains the B horizon in at least a moist state during the summer and fall, it seems unlikely that this would favor the formation of a horizon with pronounced prismatic structure. We suggest that the B horizons are in part the result of cyclic swings in climate that result in long-term lowering of the water table for clay pervection to occur and the wetting and drying to form prismatic structure. Stine (1994) reports that extensive periods of droughts lasting more than 200 years have occurred during medieval time in the Sierra Nevada. Moreover, records of periodic invasion of mountain meadows in the Sierra Nevada by lodgepole pine suggest extended dry periods occurred (Helms 1987). A forested meadow can revert back to meadow vegetation by way of fire (DeBenedetti and Parsons 1979; Kilgore and Taylor 1979). Post-fire-reduced evapotranspiration elevates the water table, and meadow vegetation returns, self-sustaining until the next extended dry period (Wood 1975). This cyclic process most likely occurred several times in the latest Holocene.

Implications

Grizzly Creek meadow is a recent geologic occurrence. Soil stratigraphy reveals that a complex interaction of wildfire, hydrology, climate, and vegetation has shaped the evolution of the meadow. Changes in the above factors can trigger a set of circumstances that lead to either valley cutting, valley filling, forest encroachment, and forest retreat. Management decisions concerning the meadow may feed back into the natural web of interacting factors and impact meadow evolution. For example, forest harvesting the surrounding uplands may raise the water table enough to alter the present plant community. Moreover, stream bank destabilization from cumulative effects of upper watershed disturbances (e.g., roads and logging) and livestock impacts may increase stream downcutting and thereby lower the meadow water table, which would favor invasion by lodgepole pine. One must be cognizant, however, that in the late Holocene, natural forces dramatically altered the character of the meadow, and these systems are dynamic. Natural disturbances may overwhelm either good or bad management of these meadow systems.

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REFERENCES

Anderson, R. S. 1990. Holocene forest development and paleoclimates within the central Sierra Nevada, California. J. Ecol. 78:470–489.

Antevs, E. 1938. Post-pluvial climatic variation in the southwest. Am. Meteorol. Soc. Bull. 19:190–193.

Bailey, S. W. 1980. Structures of layer silicates. *In* Crystal structures of clay minerals and their x-ray identification. G.W. Brindley and G. Brown (eds.). Mineralogical Society Monograph No. 5. Mineralogical Society, London, pp. 1–24.

Birkeland, P. W. 1969. Quaternary paleoclimatic implications of soil clay mineral distribution in a Sierra Nevada-Great Basin transect. J. Geol. 77:289–302.

Birkeland, P. W., and R. J. Janda. 1971. Clay mineralogy of soils developed from Quaternary deposits of the eastern Sierra Nevada, California. Geol. Soc. Am. Bull. 82:2495–2514.

Blank, R. R., and M. A. Fosberg. 1991. Effects of freezing on colloidal halloysite: Implications for temperate soils. Clays Clay Miner. 39:642–650.

Blank, R. R., T. J. Svejcar, and G. M. Riegel. 1992. Freeze-aided agglomeration of colloidal halloysite: Evidence in soils from the northern Sierra-Nevada Mountains, California. *In* Agronomy Abstracts, Annual Meeting of the Clay Minerals Society, p. 366.

Borchardt, G. 1989. Smectites. *In* Minerals in soil environments, 2nd Ed. J. B. Dixon and S. B. Weed (eds.). SSSA, Madison, WI, pp. 675–728.

Brewer, R. 1976. Fabric and mineral analysis of soils. R.E. Krieger Publ. Co., Huntington, NY.

Bremner, J. M., and C. S. Mulvaney. 1982. Nitrogen—Total. *In* Methods of soil analysis, Part 1. A. Klute (ed.). SSSA, Madison, WI, pp. 595–624.

Brinkman, R. 1970. Ferrolysis, a hydromorphic soil forming process. Geoderma 3:199–206.

Buol, S. W., and F. D. Hole. 1961. Clay skin genesis in Wisconsin soils. Soil Sci. Soc. Am. Proc. 24:377–379.

Burke, R. M., and P. W. Birkeland. 1983. Holocene glaciation in the mountain ranges of the western United States. *In* Late-Quaternary environments of the United States, Vol. 2, The Holocene. H. E. Wright, Jr. (ed.). Univ. of Minnesota Press, Minneapolis, MN, pp 3–11.

Burnett, J. L., and C. W. Jennings. 1962. Geologic map of California — Chico sheet. California Division of Mines and Geology, Sacramento, CA.

Buurman, P. N., N. Van Breeman, and S. Henstra. 1973. Recent silicifications of plant remains in acid sulphate soils. N. Jahrb. Mineral. Monatsch. 117–124.

Curry, R. R. 1969. Holocene climatic and glacial history of the central Sierra Nevada, California. *In* United States contributions to Quaternary research. S.A. Schumm and W. C. Bradley (eds.). Geol. Soc. Am., Special Paper 123, Boulder, CO, pp. 1–47.

Davis, J. O. 1978. Quaternary tephrachronology of the Lake Lahontan Area, Nevada and California. Nevada Archeological Survey, Res. Paper No 7., Univ. of Nevada, Reno, NV.

DeBenedetti, S. H. and D. J. Parsons. 1979. Natural fire in subalpine meadows: a case description from the Sierra Nevada. J. For. 77:477–479.

Duchaufour, P. 1977. Pedology. Translated by T. R. Paton. George Allen & Unwin, London.

Elmore, W., and R. L. Beschta. 1987. Riparian areas: Perception in management. Rangelands 9:260–265.

Ely, L., Y. Enzel, V. R. Baker, and D. R. Cayan. 1993. A 5000-year record of extreme floods and climate change in the southwestern United States. Science 262:410–412.

Gee, G. W., and J. W. Bauder. 1986. Particle-size analysis. *In* Methods of soil analysis, Part 1. A. Klute (ed.). SSSA, Madison, WI, pp. 383–411.

Gorbunov, N. I. 1961. Movement of colloidal and clay particles in soils. Sov. Soil Sci. 712–724.

Harris, S. 1971. Index of structure: Evaluation of a modified method of determining aggregate stability. Geoderma 6:155–162.

Haycock, N. E., and G. Pinay. 1992. Groundwater nitrate dynamics in grass and poplar vegetated riparian buffer strips during the winter. J. Environ. Qual. 22:273–278.

Helms, J. A. 1987. Invasion of *Pinus contorta* var. *murrayana* (Pinaceae) into mountain meadows at Yosemite National Park, California. Madrono 34:91–97.

Kilgore, B. M., and D. Taylor. 1979. Fire history of a Sequoia-mixed conifer forest. Ecology 60:129–142.

Kittrick, J. A. 1969. Soil minerals in the Al_2O_3-SiO_2-H_2O system and a theory of their formation. Clays Clay Miner. 17:157–167.

LaMarche, V. C. Jr. 1973. Holocene climatic variations inferred from treeline fluctuations in the White Mountains, California. Quat. Res. 3:632–660.

Lowe, D. J. 1986. Control on the rates of weathering and clay mineral genesis in airfall tephras: A review and New Zealand Case Study. *In* Rates of chemical weathering of rocks and minerals. S. M. Colman and D. P. Dethier (eds.). Academic Press, New York. pp. 265–330.

Lowrance, R. R., and R. A. Leonard. 1988. Stream nutrient dynamics on a Coastal Plain watershed. J. Environ. Qual. 17:734–740.

McKeague, J. A., and J. H. Day. 1966. Dithionite- and oxalate-extractable Fe and Al as aids in differentiating various classes of soils. Can. J. Soil Sci. 46:13–22.

McLean, E. O. 1982. Soil pH and lime requirement. *In* Methods of soil analysis, Part 2. A. Klute (ed.). SSSA, Madison, WI, pp. 199–224.

Moore, D. M., and R. C. Reynolds, Jr. 1989. X-ray diffraction and the identification and analysis of clay minerals. Oxford Univ. Press, New York.

Mortland, M. M., K. Lawton, and G. Uehara. 1956. Alteration of biotite to vermiculite by plant growth. Soil Sci. 82:477–481.

Mulholland, S. C. and G. Rapp, Jr. 1992. A morphological classification of grass silica-bodies. *In* Phytolith systematics — Emerging issues. G. Rapp, Jr. and S.C. Mulholland (eds.). Advances in archaeological and museum science, Vol. 1. Plenum Press, New York, pp. 65–89.

Nelson, D. W., and L. E. Sommers. 1982. Total carbon, organic carbon, and organic matter. *In* Methods of soil analysis, Part 2. SSSA, Madison, WI, pp. 539–579.

Phillips, W. M., and D. T. Griffen. 1981. Optical mineralogy, the nonopaque minerals. W.H. Freeman & Co., San Francisco, CA.

Reid, E. J. and G. D. Pickford. 1946. Judging mountain meadow range condition in eastern Oregon and eastern Washington. USDA Circ. No. 748.

Richards, L. A., and M. Fireman. 1943. Pressure plate apparatus for measuring moisture sorption and transmission by soils. Soil Sci 56:395–404.

Roath, L. R., and W. C. Krueger. 1982. Cattle grazing influence on a mountain riparian zone. J. Range Manage. 35:100–104.

Robert, M., and J. Berthelin. 1986. Role of biological and biochemical factors in soil mineral weathering. *In* Interactions of soil minerals with natural organics and microbes. P.M. Huang and M. Schnitzer (eds.). SSSA, Madison, WI, pp. 453–498.

Rosgen, D. L. 1994. A classification of natural rivers. Catena 22:169–199.

Sarna-Wojcicki, A. M., K. R. Lajoie, C. E. Meyer, D. P. Adam, and H. J. Rieck. 1991. Tephrachronologic correlation of upper Neogene sediments along the Pacific margin, conterminous United States. *In* The geology of North America, Vol. K-2. Quaternary nonglacial geology: Conterminous United States. R.B. Morrison (ed.) Geol. Soc. Am., Boulder, CO, pp 117–140.

Soil Survey Staff. 1984. Procedures for collecting soil samples and methods of analysis for soil survey. USDA-SCS Soil Survey. Invest. Rep. 1. U.S. Gov. Print. Office, Washington, DC.

Stine, S. 1994. Extreme and persistent drought in California and Patagonia during mediaeval time. Nature 369:546–549.

Stoiber, R. E., and S. A. Morse. 1981. Microscopic identification of crystals. Robert E. Krieger Publ. Co.,

Huntington, New York.
Thomas, J. W., C. Maser, and J. E. Rodiek. 1979. Wildlife habitats in managed rangelands — The Great Basin of southeastern Oregon. Gen. Tech. Rep. PNW-80.
Wood, S. H. 1973. Stratigraphy, chronology, and the recent erosion of mountain meadows, Sierra Nevada, California. *In* Abstracts with Programs, Annual Meeting, Geol. Soc. Am 5:868–869.
Wood, S. H. 1975. Holocene stratigraphy and chronology of mountain meadows, Sierra Nevada, California. Ph.D. thesis, California Institute of Technology, Pasadena, CA.