HOLOCENE
STRATIGRAPHY AND CHRONOLOGY
OF
MOUNTAIN MEADOWS,
SIERRA NEVADA, CALIFORNIA

Thesis by

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To Layle

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ABSTRACT

Valley-fill deposits, exposed by Twentieth-Century dissection of a number of meadows on the west slope of the southern Sierra Nevada, contain a stratigraphic record strongly affected by secular variations in watershed hydrology during the Holocene. Meadows are situated in low gradient reaches, adaquately supported by seepage water, where fine textured materials accumulate under present hydrologic conditions. Meadows do not necessarily owe their origin to glacial modification of drainage. Many meadows have formed in both glaciated and unglaciated valleys by a water table rise in valley-fill deposits.

Ground water in any meadow drainage basin is annually recharged by snowmelt. Significant evapotranspiration by meadow plants causes diurnal fluctuations of growing-season water tables on the order of 0.2 to 0.5 ft and seasonal fluctuations of 2 to 4 ft. Growing-season water-table depths are characteristically different for the two major plant communities, being usually shallower than 2 feet for meadows, and deeper than 4 ft for conifer forests. This relationship and a ground-water model are used to interpret paleohydrologic variations recorded in valley-fill stratigraphy.

Stratigraphy, radiocarbon dating, and tephrachronology indicate the following sequence in upper tributary valleys of the montane belt. Pre-Holocene cobbly alluvium rests upon bedrock. A paleosol developed upon this alluvium between 10,200 and 8700 radiocarbon years B.P., records an early post-glacial climatic interval that established forests in the present upper montane belt. The

The overlying sequence of coarse loamy materials associated with in *situ* conifer stumps indicates one or more intervals of good soil drainage and dry valley-bottom conditions between 8700 and 1200 years B.P. At some sites there is an abrupt change from forest soils to overlying wet-meadow deposits dated 2500 years at some sites and 1200 years at others, suggesting many meadows originated coincidentally with neoglaciation in the Sierras. A water-table rise of a few feet, resulting from late melting snows, could cause the change from forest to meadow conditions. Meadow deposits are composed of organic-rich, sandy-loam, topsoil layers intercalcated with sheets of well-sorted sandy gravels deposited by flood flows with recurrence intervals greater than 50 years.

A plot of upstream catchment area and valley gradient for dissected and undissected meadows indicates the geomorphic domain of unstable meadows subject to gully erosion under present hydrologic conditions on the Sierra west slope.

Two pumiceous tephra layers, widespread in meadow topsoils of the southern Sierra, are radiocarbon dated and attributed to tephra-ringed eruptive centers at opposite ends of the Mono-Inyo Crater chain of eastern California. Tephra 1, characterized by sanidine microphenocrysts and Sr content of 215 ppm, erupted 720 years B.P. Distribution of this tephra is confined to a south trending lobe extending 120 miles over the Sierra from the upper San Joaquin drainage to the Little Kern drainage. Trace element analysis of tephra 1 best match those of the tephra-ringed obsidian flow just south of Deadman Creek in the Inyo Craters. Tephra 2, characterized by a lack of microphenocrysts and Sr contents less than 20 ppm erupted from one of the northern Mono Craters eruptive centers. These two tephras appear to represent the most recent explosive eruptions of magma from this 40-km long chain of Holocene volcanoes.

CONTENTS

CHAPTER 3 -- THE STRATIGRAPHIC RECORD (Continued)

LIST OF ILLUSTRATIONS

LIST OF ILLUSTRATIONS (continued)

LIST OF ILLUSTRATIONS (continued)

LIST OF TABLES

Sugarloaf Meadow, a typical montane meadow
in Kings Canyon National Park, Sierra Nevada

footnotes for page 1

[1]Upper Sierra in this report refers to the region above 4000 feet (1220m) which roughly corresponds to the boundary between the foothill belt and the montane belt as defined by Weaver and Clements (1938). The area between 4000 and 7000 feet contains the commercial timber belt. Popular usage of the term "High Sierra" usually refers to the area above 7000 feet (2130 m) and includes the uppermost part of the montane belt, and the subalpine and alpine belt. The term "montane" describes the vegetational belt of closed conifer forest. The boundary between the montane and subalpine in the Sierra corresponds to the upper limit of red-fir dominated forest, above which lodgepole pine becomes the dominant tree in a broken forest. The term "alpine" refers to the treeless belt above conifer forests. It excludes the near timberline dwarfed conifers or krummholz; these belong in the subalpine belt. Thus the term "alpine meadow" should not be used for mountain meadows that lie below timberline. They are either subalpine or montane meadows depending on the composition and structure of the surrounding forest. Problems in world-wide usage of these terms are reviewed by Löve (1970).

[2]Meadow deposits: The deposits consist of layers of fine material (representing buried meadow surfaces), layers of alluvium, and layers of material that is largely colluvial in origin and associated with forest cover. The surficial layer, in which perennial plants are rooted, consists of admixed slope-wash sand and silt, airborne silt (including volcanic ash), and decomposed plant material. This layer has all the characteristics required of an "A" horizon of a soil profile, except that there has been no visible alteration of underlying or intercalcated alluvial layers. In this report, the surficial layer of meadows will be termed "topsoil" to avoid confusion with the term "soil," which in common geologic usage means a pedogenic soil. Materials accumulated at the base of forested slopes, is termed "soil," as these materials usually have incipient profile development; however, in recently emplaced materials, distinctions between soil, colluvium, and slope-wash alluvium cannot usually be made. The only true pedogenic soil described in the section is a buried profile near the base of the deposits.

[3]Holocene: time-stratigraphic use of the term "Holocene," in this report, refers to an epoch extending from the present back to 10,000 radiocarbon years (B.P.) as suggested by Hopkins (1975).

Organization of report

This introductory chapter describes the method of study and the physical setting of the principal meadow belt of the west slope. Chapter Two is a survey of surficial aspects of meadows, the present vegetation, and the groundwater and surface water hydrology. In Chapter Three detailed descriptions of stratigraphic sequences beneath montane meadows are followed by a discussion of Tuolumne Meadows, the only sub-alpine meadows studied in detail. These descriptions provide the basis for interpretations of the stratigraphy and its implications concerning Holocene climatological variation and the origin and maintenance of mountain meadows. Chapter Four deals with the character, age, and origin of volcanic ash layers found in meadow and other surficial deposits of the Sierra and outlines the Holocene eruptive history of the Mono-Inyo Crater volcanic chain.

Methods of study

Information on dissected meadows was obtained from Sumner and Leonard (1948), Sharsmith (1959, and 1962, and personal communication, 1972) and R. Flynn, U.S. Forest Service, Fresno (personal communication, 1972). Others were located by examining meadows identified on 1:62,500 topographic maps sheets. Dissected meadows are listed in Appendix A.

Stratigraphy in gully walls was charted on 1:48 vertical, and 1:1200 horizontal scale. Information was obtained to depths exceeding 10 feet below the water table or in undissected meadows by use of a bucket auger and a soil coring tool fashioned from 1.5" steel tubing. Control was maintained by hand level, plane table alidade, and stadia rod, and cloth tape. Topographic maps were made of three meadows on a scale of 1:2400 and 5-foot contour interval.

Materials were described in the field in a manner similar to that recommended by Birkeland (1974). Suspected volcanic ash layers were usually distinguishable from fine white sand or diatom layers by rubbing the material between the fingers. Angular sharp glass shards of volcanic ash have a characteristic rough feel compared to sand and diatoms. Very thin or questionable layers were sampled and later verified by microscopic study. The genus of stumps and logs in the stratigraphic section could usually be identified by field examination under hand-lens of transverse wood sections smoothed with a razor blade.

Materials for radiocarbon-age determination are so abundant that only pure wood and charcoal were used, thus avoiding the uncertainty occassioned by penetration of recent roots into other organic matter and peat. Wood up to 10,000 years old is remarkably well preserved in meadow deposits. It was usually possible to count the tree rings and sample a known tree-ring interval of less than 20 years. Samples were examined for visible contaminants, and parts in contact with younger rootlets were cut out. Materials were oven dried at 100° C and submitted to the Geochronology Laboratory of the University of Georgia. The laboratory treated the materials to remove soil carbonate or soluble soil humates prior to synthesis of benzene for liquid scintillation counting. Radiocarbon ages as listed in the appendix and have been published in *Radiocarbon* by Brandau and Nokes (1974).

Radiocarbon ages are reported as years before present (B.P.), which is the number of years prior to 1950 A.D. determined from a carbon-14 half-life of 5570 years. To convert ages in this report to the presently accepted half-life of 5730±30, ages should be multiplied by a factor of 1.03. Work by Ferguson (1968) and several radiocarbon laboratories calibrates radiocarbon ages to calendar ages (See Olsson, 1970 or Ralph, 1974 for a review). In the range 3000 to 7000 calendar years, radiocarbon ages are approximately 10 percent younger than corresponding calendar ages. The reader is referred to curves published by Suess (1970) or Ralph (1974) to convert from radiocarbon to calendar age. Beyond 6800 years B.P. radiocarbon ages remain uncalibrated, although the varve chronology established by Stuiver (1970) is often used to correct radiocarbon ages beyond the range of bristlecone-pine chronology.

Geological setting

The Sierra Nevada is a great west-tilted fault block composed primarily of Mesozoic batholithic rocks. The present west slope rises at an average slope of 4° from near sea level at the east side of the Central Valley to heights exceeding 12,000 feet along the crest. The transverse profile of the range is strongly assymetric with a precipitous east slope descending on an average of 15° to adjacent eastside basins most of which have floors 4000-5000 feet above sea level. The principal belt of meadows (figure 1-1), generally over 20 miles wide, lies between 5000 and 9500 feet elevation on the west slope.

The batholithic rocks are largely coarse grained granodiorite and quartz monzonite with large roof pendants of metamorphosed Paleozoic and Mesozoic rocks along the crest

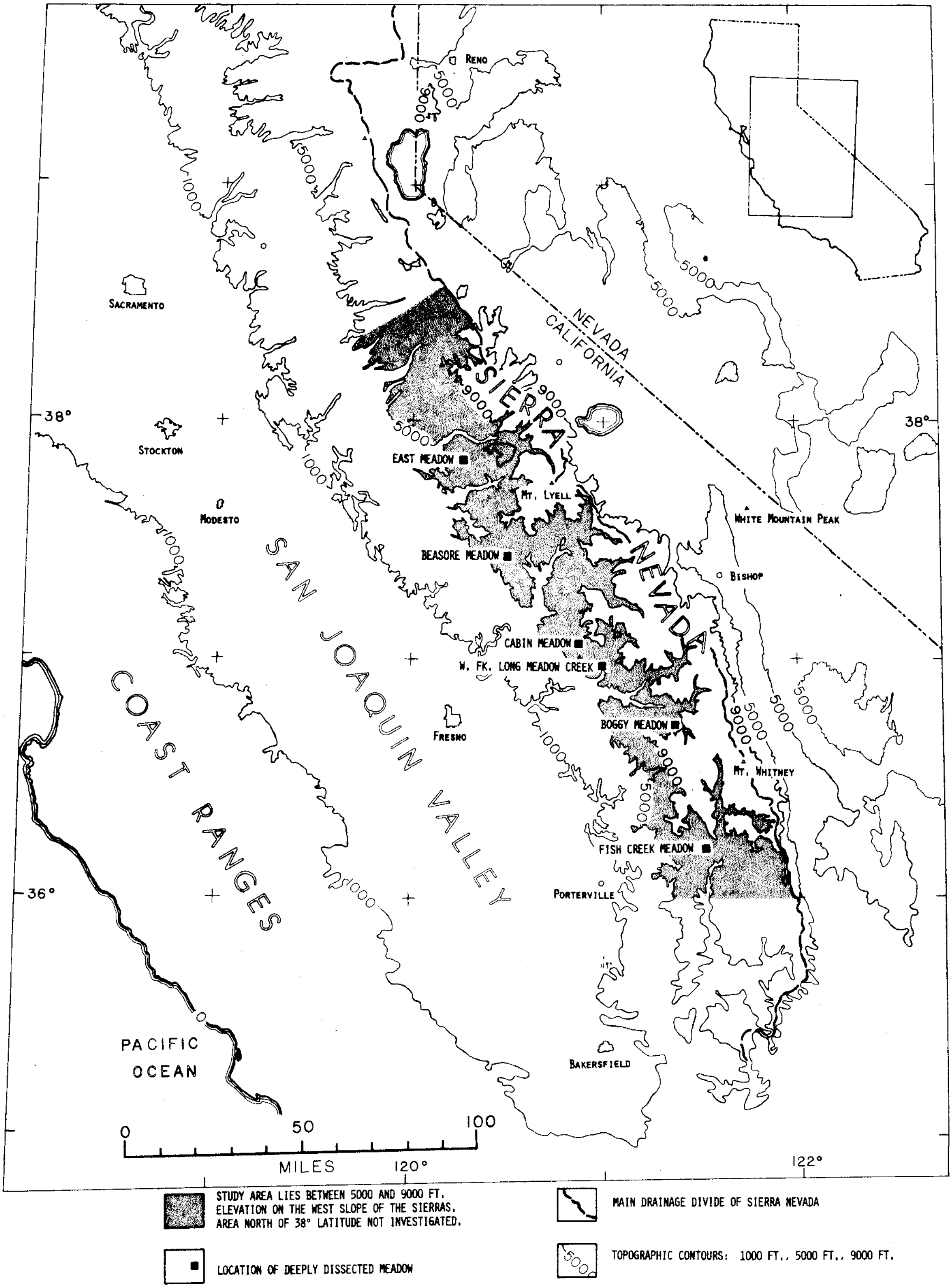


Figure 1-1. Location of the principal west-slope meadow belt, southern Sierra Nevada, showing locations of deeply dissected meadows.

and at several places on the West slope. Tertiary volcanics cover part of the batholith north of the Tuolumne River. Scattered remnants of late Tertiary and Quaternary volcanics crop out in the valleys and uplands of the southern Sierra (Bateman and Wahrhaftig, 1966).

Christiansen (1966) concludes the present elevation and tilted west slope of the Sierra are the result of broad crustal upwarping involving also the Basin and Range province to the east during a period between 9 and 3 million years ago. Available evidence suggests major downfaulting that created the eastern escarpment did not begin until the Pleistocene.

The high Sierra was extensively glaciated during the Pleistocene. Major canyons harboured large valley glaciers that descended to elevations below 5000 feet. Little of the country below 7500 feet was glaciated except within major canyons.

The slope of the west-central Sierra rises eastward in five or more steps, each step being a few hundred to a few thousand feet high. Steps are irregular but generally follow contours of the western slope for up to 10 miles. Treads are from less than one mile to more than 5 miles wide, and some have eastward slopes (Wahrhaftig, 1965). Valleys in which most of the montane meadows lie are on these treads. Except along glaciated high divides of the west-slope uplands, hills are generally rounded and valleys have wide open V shapes. The major west-draining rivers incise this upland surface. Prominent knickpoints of most gorges lie within 5 to 10 miles of major drainage divides. Between knickpoint and divide, the rivers flow in broad open valleys on the upland surface.

Matthes (1930) proposed that upland surfaces of the Sierra are the result of repeated cycles of uplift and erosion during the Tertiary, whereas Wahrhaftig (1965) has suggested that the steps and treads of the west-slope are a normal product of hillslope erosion of a granitic terrain and probably the result of a single uplift, which, according to Christiansen (1966), occurred mostly in the Pliocene.

Rates of landscape evolution can be estimated from Jahnda's (1966) reported denudation rates in the upper Sierra of less than 0.08 feet/1000 years as judged from reservoir sedimentation studies. From volumes of Pleistocene sediment in the San Joaquin Valley, he suggests that denudation rates during late Wisconsin time may have been an order of magnitude greater. Canyon deepening of 0.3 feet/1000 years both by glacial and fluvial erosion is indicated by 300 feet of downcutting of the upper San Joaquin River through one-million-year-old basaltic-andesite flows (Rinehart and Huber, 1965) and underlying granitic rock. Thus, the current physiography involves at least several hundred feet of

Pleistocene erosion. During the Holocene, it is unlikely that bedrock physiography of the west-slope has been modified by more than a few feet. Hence, the low bedrock-gradients of valley floors on which most of the meadows lie were presumably well established before the Holocene, and are probably the result of erosion over the preceding several hundred thousand years.

CHAPTER 2

MEADOW DEPOSITS, SOILS, VEGETATION, AND HYDROLOGY

Physical characteristics of meadow settings

Meadows are characterized by two fundamental physical conditions: 1) a shallow water table that rarely exceeds two feet in depth at mid-summer, and 2) surficial material that is fine-textured and richly organic (classified by Bennett (1965) as largely sandy loams). The diversity of sites upon which the meadow plant community is found is described by Sharsmith (1959) who writes:

> The back country meadows occur in various situations: on forested or brushy, sometimes fairly steep slopes, in swales or large level or gently inclined spaces more or less surrounded by often heavy forest, on flats in valley or canyon bottoms, in level, filled-in lake basins, and on open, subalpine or alpine slopes and benches. In the zone of forest they range from minute representations to large clean expanses of sixty or more acres, while at open, higher altitudes, they may extend brokenly over almost indefinite distances.

Most montane and subalpine meadows described in the present study are situated on gently sloping flats on valley floors surrounded by conifer forest.

Geomorphic domain of meadows

Conditions favoring shallow water tables and accumulation of fine-textured materials are met where relatively impervious bedrock floor of a valley has a gentle gradient. Size of the drainage basin upstream from the meadow is also a factor, since large basins tend to yield high discharges which carry away fine materials. Nonetheless, the basin must be large enough to provide seepage water adequate to maintain a shallow water table during the growing season. Presumably there is a ratio of drainage basin area to meadow area below which the amount of seepage water supplied to the site will be insufficient to maintain a high water table. The lowest ratio measured is 2.6. Most ratio values lie between 5 and 25, but range up to several hundred for meadows with large through-flowing streams (Tuolumne Meadows for example).

More critical to the existence of a meadow appears to be the relationship between drainage area and surface slope of the meadow. A plot of these parameters (figure 2-1), displays a domain in which no meadows are know to exist in the part of the Sierra studied. Steep valley-bottom gradients and large upstream catchment areas are characteristic of this domain. Valley reaches in this domain have water tables more than several feet deep, and they are forested. At steeper slopes or with larger catchments, the stream regime through the valley prevents accumulation of alluvial and slope-wash fines, and water flows upon bedrock.

A domain of "unstable meadows" is identified on the plot by a group of deeply incised meadows. It is thought that most incised meadows built up steeper slopes than allowed by the drainage-basin size under the protection of the tough sod of the meadow community. Use of these meadows by livestock may have weakened the protective sod and allowed erosion to take its course. At lower slopes and under smaller drainage basins is a domain of meadows considered stable at the present time. Here fine materials accumulate, and even under heavy use by livestock these meadows do not appear susceptable to the deep dissection observed in the "unstable domain." Only one of the meadows studied appears exceptional as shown by two points in the stable domain (figure 2-1). This is Long Meadow on the North Fork of the Kings River. Water has been channelled along well used stock trails and is presently cutting steep-sided gullies six feet deep at the upper end of the meadow. Gullies die out downstream, and there does not appear to be any threat to meadow stability at the lower end.

Meadow Topsoil

Meadows are depositional sites, and they are underlain by stratified alluvial and organic accumulations. The topsoil is 0.5 to 3 feet thick and commonly rests upon a layer of well sorted alluvial sand that may be a few inches to one foot thick. The uppermost 0.3 ft. of topsoil is a tough root-bound sod layer, gradually transitional to an underlying dark-brown sandy loam. Below the root-bound sod, roots are fewer and finer and rarely extend deeper than one foot from the surface.

Properties of 50 topsoil samples of 10 selected meadows determined by Bennett (1965) probably encompass the range of properties encountered in montane and subalpine meadows. Organic content, reported as per cent weight-loss after oxidization by hydrogen

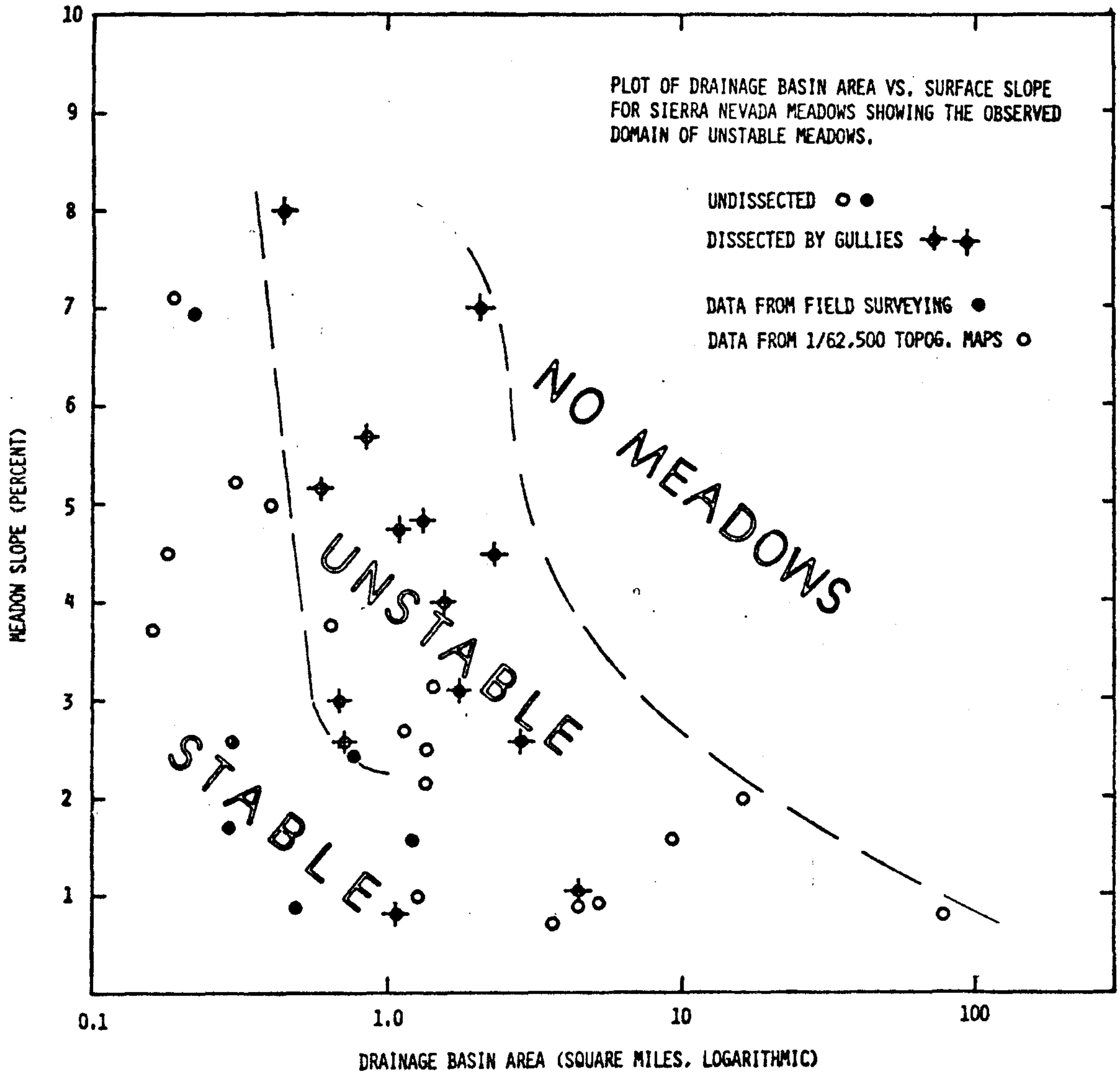


Figure 2-1. A plot of meadow-surface slope versus drainage basin area above meadow for west-slope Sierra Nevada meadows.

peroxide or incineration of samples dried at 105° C, ranges upward from 4 per cent to nearly 100 per cent in pure sedge peat, but most samples contain about 5 per cent. Mineral constituents are fine and the topsoils are classified largely as sandy loams. Textural clay is less than 6 per cent of the inorganic fraction, and mineral clay is probably absent.

Drained topsoil samples commonly have a moisture content of 40 per cent by weight, ranging up to 80 per cent for peat. The colors are dark brown to black, reflecting mainly the color of moist organic matter. Virtually all meadow topsoils examined in the Sierra are acidic, pH values ranging 3.7 to 6.7 with a value of about 5.0 most common (Bennet (1965). The alluvial sand beneath the topsoil shows no profile development. Buried materials deeper in the section show similar alternations of dark sandy-loam on alluvial layers, none of which have developed true pedogenic horizons. These alternations of a thin topsoil on sandy alluvium extend up to depths of 15 feet. Still deeper are older sequences of materials described in Chapter 3. Total thickness of valley fill beneath meadows in montane valleys does not usually exceed 30 feet.

Soil moisture regime has only recently become an accepted parameter for classifying soils (Smith, 1973), yet it is the most important single property of meadow topsoils. In the United States system of soil[1] classification (Soil Survey Staff, 1973) the moisture regime of meadow topsoils ranges from perudic (moisture control section[2] at or very near field capacity throughout the year) to udic (moisture control section not dry for more than 90 days out of the year, and soil not saturated with oxygen depleted water). The combination of shallow water tables and fine materials produce the udic to perudic conditions in meadows. As the water table declines in the latter part of the summer, the fine topsoils can conduct water to the shallow-rooted meadow plants by capillary rise, for the height of capillary rise in very fine sand (0.1 to 0.05mm) can exceed two feet in 24 hours (van't Woudt and Hagan, 1957). In coarse sand, the rise is only a few inches. Fine silt and clay-size materials can conduct water to greater heights, but the rate is too slow (figure 2-2) to support the meadow plant community. The mean particle size of meadow topsoils falls near 0.1 mm which gives the optimal 24-hour rate of capillary rise. This mechanism of drawing water from depths of a few feet provides the meadow community with an adequate soil moisture regime during the largely precipitation-free summers. It also explains why sandy gravel bars deposited on meadows do not rapidly become vegetated. The water table must either rise into the gravel bar, or silt must footnotes for page 10

[1]Soil: In discussions involving standardized terminology for soil properties defined by the Soil Survey Staff (1973) and the Soil Science Society of America (1973), it is necessary to use the term "soil" for any material at the earth's surface capable of supporting plants. Exception is made in this chapter only to the geologic use of the term "soil." Where ever possible, plant-supporting surficial materials that are largely depositional in origin are referred to as "topsoils."

[2]Moisture control section has an upper boundary at the depth to which a dry soil is moistened by 2.5 cm of water in 24 hours and a lower boundary at the depth to which the soil is moistened by 7.5 cm of water in 48 hours. Dry soil has water held at tensions greater than, or equal to 15 bars. Moist soil has water held at tensions less than 15 bars (Soil Survey Staff, 1973).

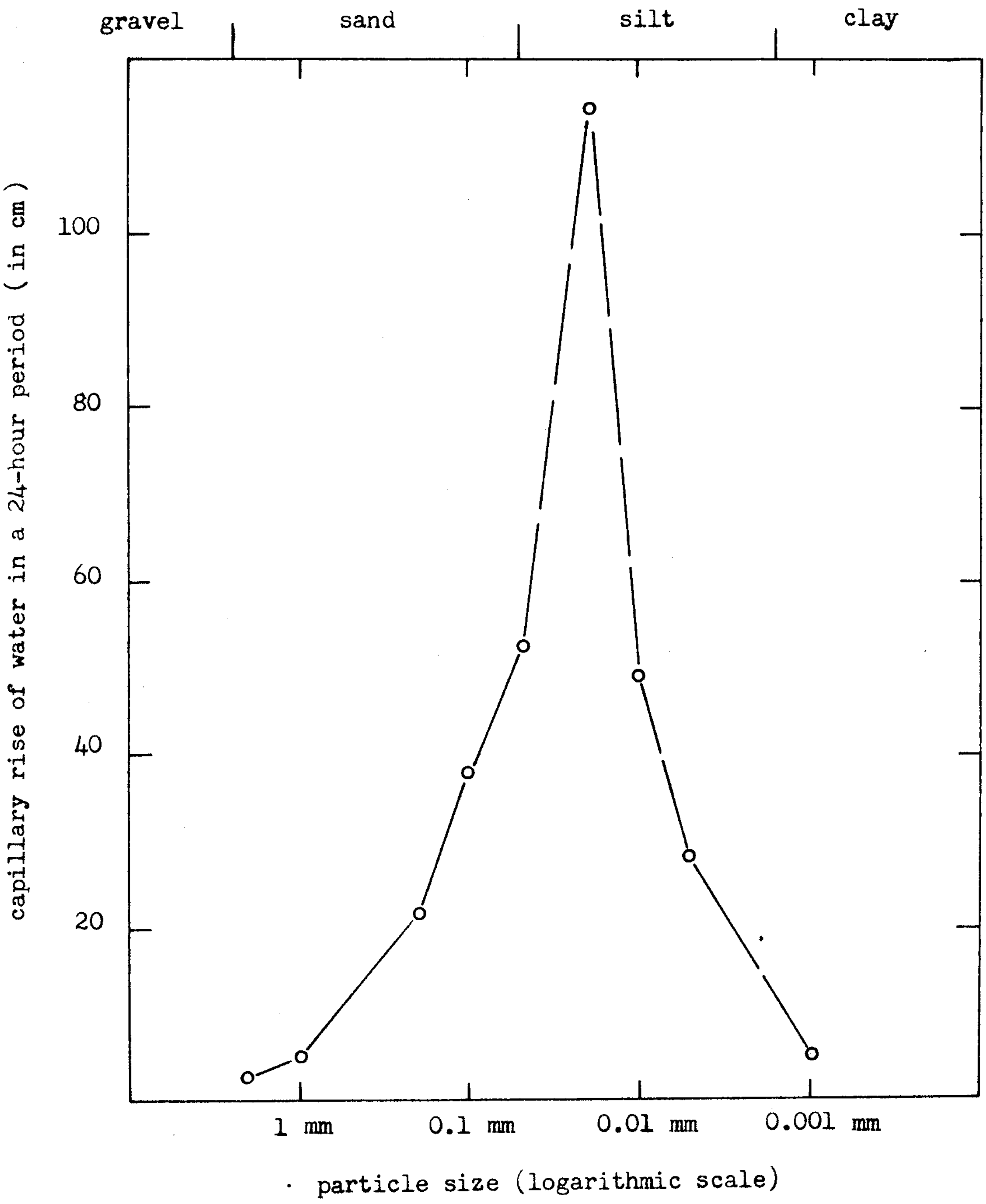


Figure 2-2. Height of capillary rise observed 24 hours after application of a water table to columns of gravel, sand, and soil at 17°C. After Atterberg (1908), from van't Woudt and Hagan (1957)

accumulate in it in order to provide a suitable moisture regime in summer for the shallow-rooted meadow plants.

In terms of the U.S. system of soil taxonomy (Soil Survey Staff, 1973) topsoils of mountain meadows would be classified as *udic* and *perudic cryofluvents* of the *Entisol* order, although in the wettest parts of the meadow, topsoils with organic matter content exceeding 20 per cent would be called *fluvaquentic borofibrists* or *borohemists* of the *Histosol* order.

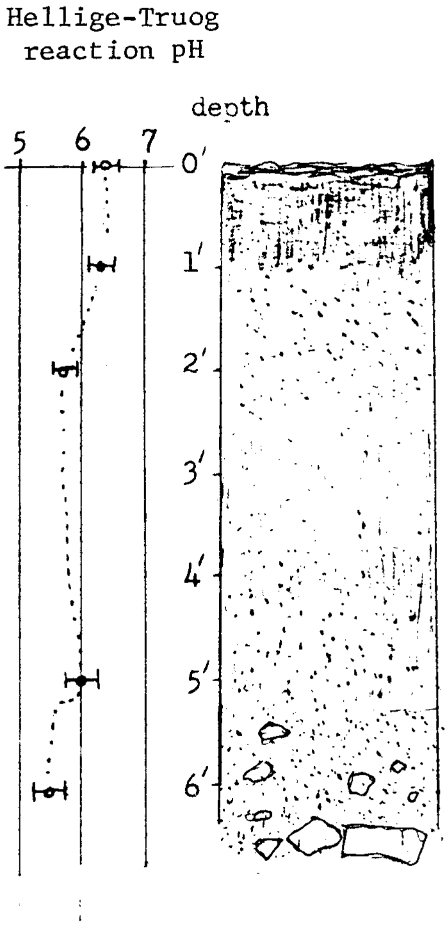
Forest Soils

Soils of the surrounding forest differ in texture, organic content, and moisture regime from the meadow soils. Forest soils at the foot of slopes have both alluvial and colluvial components, but higher on the slopes they may develop residually upon bedrock. Most forest soils of the montane and subalpine belt have a generally weak profile development characterized by a thin surface duff, a slightly organic "A" horizon with scattered charcoal and rotted wood fragments, and judging from color, probably less than 2 per cent organic, a weak color "B" horizon (yellowish-brown), and a "C" horizon composed largely of residual or transported grus (figure 2-3). Horizon boundaries are transitional and indistinct. Analyses of 208 forest soil samples by Willen(1965) show that they are coarse textured with 25 per cent or more of the grains larger tha 2 mm. The remaining matrix is loamy sand, containing less than 6 per cent clay sized material. Colluvial and slope-wash layers are unsorted and crudely stratified, water tables are commonly more than 4 feet deep, and slopes range upwards from 3 to 10 percent. The topographic relationship of forest soil and vegetation to the meadow is shown in figure 2-4. The transition between soil types may occur over a distance of 30 feet.

Sierra forest soils have been regarded in the past as regosols or andosols (Jahnda, 1965 and Huntington, 1962). Under the present United States system of soil classification (Soil Survey, 1973) they would be classed as members of the *Inceptisol* order.

Vegetation

Although sedge, largely of the genus *Carex*, is the most conspicuous component of montane meadow vegetation, species of rushes, herbs, grasses and dwarfed shrubs are important and in some meadows are more abundant than sedge. Representative transect



O: surface duff (needles and branches).

A: sandy loam with admixed needle litter, charcoal and wood fragments.

B: sandy loam, dark yellowish brown (10 YR 4/4), friable and loose.

loamy sand, dark yellowish brown (10YR 3/4), friable and loose.

C: loamy sand, dark brown (10YR 3/3) friable and loose.

C: medium grained sand, light yellowish brown (10 YR 6/4), slightly sticky.

Figure 2-3. Profile description of forest soil on the edge of Exchequer Meadow, Sierra National Forest.

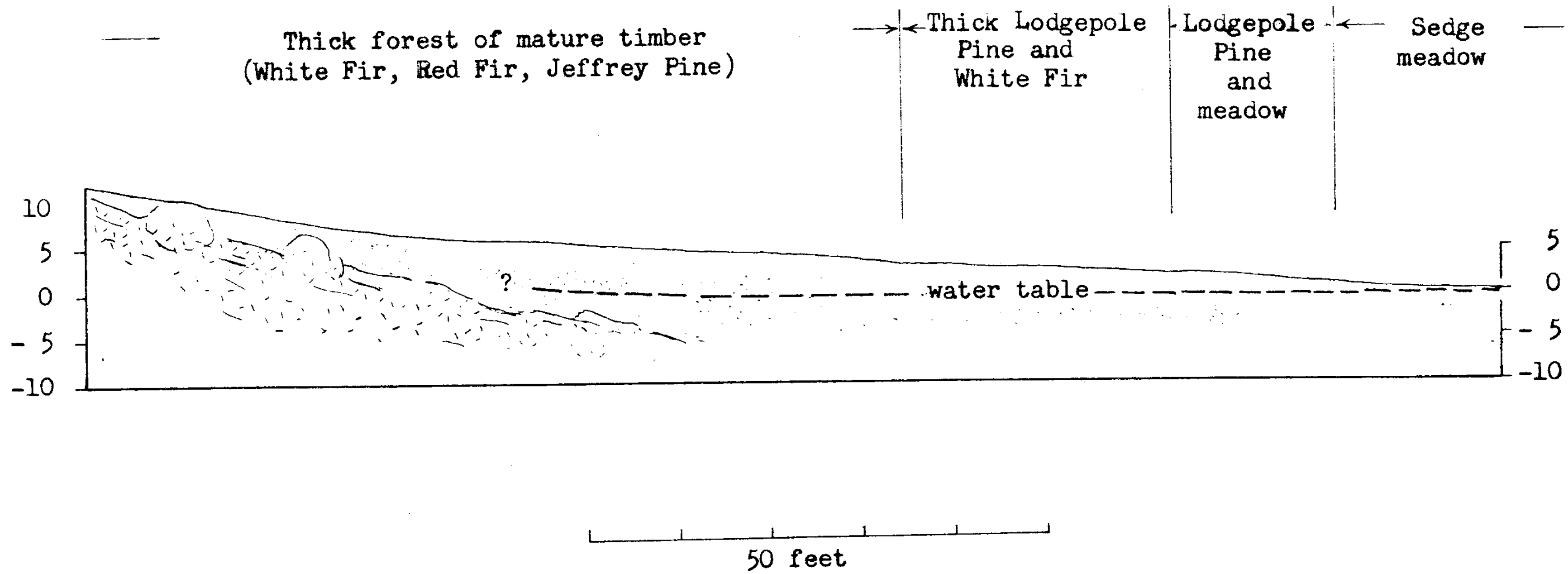


Figure 2-4. Cross-section showing relationship of the forest - meadow ecotone to the early-summer water table based on data from auger holes, south side of Exchequer Meadow, Sierra National Forest, California.

TABLE 2-1: Vegetation Transect surveys in montane and sub-alpine meadows, west-slope, Sierra Nevada (percentage of ground covered by each species)

	Upper meadows, Paradise Valley, KCNP elev. 6620 ft. (Strand, 1972)		Jackass Meadow, Sierra National Forest, elev. 6500-7000 ft. (Sanderson, 1967)		Junction Meadow KCNP elev. 8160 ft. (Strand, 1972)		Exchequer Meadow, Sierra National Forest elev. 7200 ft. (this study)	
	Thin leaf sedges	27.7	*C. Nebraskensis*	11	Fine leaf sedge	12.2	Thin leaf sedge	31.1
	Medium leaf sedges	9.9	other *Carex*	12	Broad leaf sedge	6.3	Broad leaf sedge	12.2
			probably *Scirpus* and *Eleocharis*	40	Baltic rush	12.2	Spike rush	5.6
TOTAL SEDGE & RUSH		37.6		63		30.6		48.9
	Perennial grasses	27.4	*Muhlenbergia filiformis*	21	Perennial grasses	20.1		
			Danthonia california	2				
			Agrostis idahoensis	1				
TOTAL GRASSES		27.4		24		20.1		- 0 -
	Aster	3.1	Legumes, probably *Lupinus spp.* and *Trifolium spp.*	3	flowering herbs and forbs	7.8	Herbs (monkey flower and others)	42.3
	non-plant	8.7	Other	10	non-plant area	7.4	mosses	4.6
	other	23.2	*Dodecatheon jeffreyi* and other non-palatable species not counted				Other	3.6
TOTAL MISC.		35.0		13		15.2		50.5
TOTAL		100.0		100		100.0		100.0

	Charlotte Meadow KCNP elev. 9200 ft. (Strand, 1972)		Vidette Meadow KCNP elev. 9520 ft. (Strand, 1972)		Gaylor Lakes Basin Yosemite Natl. Park elev. 10,000 (Klikoff, 1965)		Gaylor Lakes Basin Yosemite Natl. Park elev. 10,000 ft. (Klikoff, 1965)	
	Sedge	3.4	Thin leaf sedge	11.0	*Scirpus clementis*	5.4	sedge	1.0
			broadleaf sedge	5.9	*Carex spp.*	1.7		
TOTAL SEDGE & RUSH						7.1		1.0
	Wet meadow complex of *Heleocharis breweri* (spiked rush) and *Calamagrostis pauciflora* (grass)	54.7	shorthair grasses and rushes	22.6	*Calamogrostis*	33.4	*Poa*	37.0
			perennial grasses and spiked rush	37.8	*Deschapsia caesppitosa*	2.2	*Calamogrestis*	9.0
TOTAL GRASSES						35.6		46.0
	Kalmia Polifolia	7.7	aster	10.2	*Aster*	23.5	*Antennaria*	23.0
	Vaccinium nivictum	7.1	*Vaccinium nivictum*	2.7	*Vaccinium nivictum*	15.8	*Potentilla*	9.0
	gentian and aster	5.3	non-plant	5.4	*Kalmia*	5.0	*Saxifrage*	up to 18.0
	non-plant	17.5	other (11 species)	4.2	dead plants	9.0	*Draba*	1.0
					other	4.0	*Lupinus*	1.0
							other	3.0
							bare ground	18.0
TOTAL MISC.		37.6		22.5		57.3		53.0
TOTAL		95.7		99.8		100.0		100.0

data of species composition is hard to obtain because the meadow is an intricate complex of many small communities associated with slightly different soil moisture regimes. Investigators adopt different means for grouping species of uncertain identity, hence few reported surveys are easily compared. However, available surveys provide a general view of meadow vegetation (table 2-1). Grasses tend to dominate the cover of upper subalpine meadows, and sedges and rushes prevail in the lower subalpine and montane meadows. In lower meadows, considerable species diversity is reported, with upwards of 30 species along a transect (Sanderson, 1967, and Strand, 1972).

If ungrazed by stock, mid-summer foliage of moist montane meadows grows knee-high, but in subalpine meadows foliage rarely exceeds one foot. Heavy stock grazing has altered species composition at many meadows (Sumner,1940,Sharsmith,1959,Storer and Usinger,1963 Bennett, 1965, and Strand, 1972). Increase in percentage of barren ground and cover by non-palatable species are cited as typical evidence of meadow deterioration by grazing.

Plant ecology

Mountain meadow communities occupy moist sites underlain by shallow water tables. Meadows are often spoken of as grasslands because they constitute a mosaic of open communities within the forests. True grasslands, however, occupy relatively dry sites. Groves of trees associated with grasslands in the Central Valley of California seek the north facing slopes or riparian sites to avoid summer drought. Grasslands on dry hillsides and alluvial plains have a growing season limited to the wet season of late winter and early spring (Gleason and Cronquist, 1964). Just the opposite occurs with mountain meadows and the continuous conifer forest. Forest take the dry, drained, sites and meadows are established on moist sites.

The mixed conifer forest of the west Sierra slope grows on a deep, aerated, drained soil with a mid-growing season water table deeper than 4 feet. Soil volume available to the open system of coarse tree roots would be restricted by shallower water Table conditions. The trees would be unable to extend sufficient lateral roots to provide needed root area for gathering nutrient mineral requirements. It is known some trees can tolerate several weeks of flooding during dormancy but can be seriously injured by only one day of flooding in the growing season (Kramer, 1969). Waterlogging in winter or early spring may not effect trees, but if soils do not drain in late spring or early summer tree growth may be inhibited.

Unforested tracts are meadows, dry rocky slopes with shallow soil, or deep soiled sparsely vegetated flats. These flats, which appear favorable to forestation, may have growing-season water tables too shallow for growth of conifers, but too deep for shallow rooted meadow plants, or perhaps lack mycorrhiza soil fungi (Wilde 1968, and Went and Stark, 1968) necessary for the growth of most coniferous trees.

Favorable conditions for plant growth characteristic of mountain meadow situations are fine-textured organic topsoils, high soil-moisture content, open sunny sites, and warm clear summers of the Sierra.

Two unfavorable conditions too which meadow plants have adapted are the relatively short 110-day frost free growing season of the montane belt, and the shallow ground-water table which severely limits the thickness of topsoil suitable for the physiological functioning of plant roots.

Sedges and perennial herbs adapt to the short growing season by their ability to burst forth a flush of foliage as soon as snow disappears from the meadow. Growth dynamics of these plants is related to their ability to store significant carbohydrate reserves in their below-ground biomass (discussed in a later section).

Meadow plants cope with shallow water-table conditions, by concentrating their roots in the upper few inches of topsoil. The root system is a dense mass of fine filamentous roots providing large root area per unit volume. During periods when topsoil is water-saturated and poorly aerated, specialized physiological processes allowing respiration under anaerobic conditions, (McManmon and Crawford, 1971) and cell structure allowing flow of oxygen from above ground to roots (Arber, 1920, quoted in Moore and Bellamy, 1974) may allow some meadow sedges, rushes, and herbs to continue normal growth.

Aeration of waterlogged topsoil is limited by the diffusion rate of oxygen in water which is 7300 times less than the rate in air (Kramer, 1969), but permeable topsoils may also be aerated by a high flux of modestly aerated groundwater. Water saturated topsoils of meadows must be moderately aerated during much of the year for blue-gray soil coloration of gley horizons is not encountered. This indicates oxygen concentrations above 0.2ppm and oxidation potentials above +300 millivolts in pH 5 soils (Daniels and others, 1973). Kramer (1969) suggests soil atmospheres above 0.1 oxygen partial pressure (equilibrium concentration in soil water would be 4 ppm) are adequate for growth of most plants. Fully aerated water at 16° C in equilibrium with the atmosphere contains 10 ppm oxygen. Presumably, floras adapted to watersaturated soil conditions are tolerant of oxygen concentrations in the range 0.2 to 4 ppm.

Water-table depth controls the position of the transition zone (ecotone) between meadow and forest. This depth fluctuates during the growing season, but a profile of the ecotone (figure 2-4) indicates that the water table beneath forests must be deeper than 4 feet by early July. Meadow water tables are near the surface at this time, but by the end of August they may drop to depths of 2 to 4 feet.

Life cycle and biomass production of meadow plants

Growth habits of meadow plants allow the community to flourish in the relatively short growing season between snowmelt and autumn frosts. Understanding of seasonal changes within the meadow community and of the life cycle of individual plant species leads to a better comprehension of soil formation, maintenance of the meadow, and suceptability of the meadow surface to damage through improper use. Within a week after snow disappears from any part of a meadow, sedges vigorously send forth green shoots and snowfree areas are soon a verdant sward of sedge blades. Vigorous growth of grasses and herbs seems to be delayed until mid-summer, and meadows reach their maximum foliage developemnt by mid-July. The foliage begins to brown in August, and by the first of September, meadows are bronze-yellow fields. Hydrologic data presented in a later section show a decline in transpiration rates by mid-August suggesting that senescence rather than declining soil moisture or autumn frosts is responsible for browning the foliage. Exception to this occurs during dry summers following abnormally low winter snowpacks when wilting and browning occurs on less favorable moisture sites at meadow edges.

Perennial sedges and rushes dominate most montane meadow communities. These plants maintain themselves, or extend their territory largely through growth of highly developed, horizontal, underground stems called rhizomes, from which new shoots sprout. Although these plants produce flowering stems at maturity, it is unlikely that reproduction by seed is a major factor in extending or maintaining the meadow.

The life history of sedge, though little studied in the Sierra, has been investigated on a sloping montane fen in the Alberta Rocky Mountains (Bernard, 1974). These studies concern the broadleafed sedge, *Carex rostrata*, a species common on wet montane meadows in the Sierra. Two remarkable features of the sedge emerge from these studies. These plants store carbohydrate reserves during the late summer in their below-ground biomass (roots and rhizomes) which allows the plants to burst forth vigorously as soon as snow cover is removed in the spring, a capability originally reported by Mooney and Billings (1960) from studies of alpine meadow and fell-field perennial herbs. The second feature found in the broadleafed sedge, is that the plant may have two different longevities (18 months or 24 months) depending upon whether shoots emerge from the rhizomes in late summer or in the following early spring. New shoots of the sedge leaf-out the first summer and die back in the fall. The following spring, new blades grow within the sheath of the previous summers foliage and the flowering stems emerge and mature by mid-summer. In late summer, the foliage browns and dies back. New shoots will emerge from the extended rhizome of the matured plant either in the fall or early spring. Shoot growth occurs while the plant is snow covered, as it prepares for the flush of growth when the snow melts. This life history demonstrates the importance of late fall and winter plant behavior to the vigor of the meadow in the spring. If the meadows are extensively grazed and trampled in July and August, photosynthetic production required to produce reserves of carbohydrate would be reduced, thus inhibiting spring growth.

Late summer standing meadow vegetation reported in several studies in the Sierra suggest an annual productivity for the above-ground biomass of 100 to 200 gm/m^2 (table 2-3). These values are obtained by clipping all active vegetation 0.5 inches above the ground on averaged 1-ft^2 plots and weighing the air dried material. They are comparable with sedge dominated wet meadows in the Medicine Bow Mountains, Wyoming but considerably less than the montane fen in Alberta or the sedge wetland of Minnesota.

From Bernard's (1974) measurements of above-and-below-ground biomasses in the sedge wetland in Minnesota, a relatively complete picture can be obtained of the annual budget of organic material in a sedge community. In the summer, 845 gms/m^2 are contained above the ground, and 185 gms/m^2 below the ground. In the winter, 114 gms/m^2 are above the ground and 328 gms/m^2 are below the ground. The increase in below-ground biomass of 140 gms/m^2 is attributed to the transfer of carbohydrate to the roots and rhizomes where it is stored as winter reserves.

It may be theoretically possible to estimate the annual budget of a mountain meadow if the budget for the sedge wetland (852 gms/m^2 above-ground biomass) can be directly

Table 2 - 3

LATE SUMMER ABOVE-GROUND BIOMASS OF SIERRA NEVADA MEADOWS

Location	Plant community	Above ground biomass	Reference
Marshall Meadow, Jackass Meadow, Sierra National Forest, 6500 - 7000 foot elev. (upper montane)	moist meadow (rush, sedge, grass)	94 - 116 gm/m^2	Sanderson (1967)
Vidette Meadow, Kings Canyon National Park, 9520-foot elev. (subalpine)	moist short-hair grass meadow (*Calamagrostis*)	115 gm/m^2	Ratliff (1972)
	open meadow ?	228 gm/m2	
Gaylor Lakes Basin, Yosemite National Park, 10,000 foot elevation. (timberline)	wet short-hair grass (*Calamagrostis*) meadow	180 gm/m^2	Klikoff (1965)

ABOVE GROUND BIOMASS OF MEADOWS AND WETLANDS AT OTHER GEOGRAPHICAL LOCALITIES

montane fen, Rocky Mountains, Alberta, elev. 4770 feet	wet fen *Carex rostrata*	515 gm/m^2	Gorham and Somers (1973)
alpine meadow, Medicine Bow Mountains, Wyoming,	wet meadow	114 gm/m^2	Billings and Bliss (1959)

scaled to a meadow in the Sierra (100gms/m^2 above-ground biomass). Knowledge of the organic matter budget would provide insight into two vital processes pertinent to maintenance of meadows subject to foliage loss through trampling or grazing: the translocation of carbohydrate to the roots in late summer, and the net annual input of organic matter into the topsoil. Scaling Bernard's data by a simple factor of 100/852 implies an organic matter budget shown in figure 2-5. During late summer an average of 17 gms/m^2 is translocated to the underground reserves. Actual annual transfer may be somewhat greater as no allowance is made for lost storage in roots and rhizomes that died during the year. This loss should amount to less than one-half the summer underground biomass for perennial plants having two-year longevities, or about 10 gms/m^2.

In the above-surface biomass about 70 gms/m^2 will be shed as brown stems and leaves at the end of summer. Part of this dead foliage as well as the dead root material will decompose and be lost as gaseous CO_2. The remainder will become soil organic matter. Rate of plant matter decomposition in meadow soil is not accurately known, but experimental work of Waksman and Gerretsen (1931) allow a rough estimation. They found that 55 per cent (by dry weight) of fresh oat straw decomposes into CO_2 in 273 days leaving behind a residue of loose brown peat and humus. Conditions were kept at 18^o C and moisture at 80 per cent field capacity, conditions not unlike those in montane meadows. A reasonable estimate is that 45 per cent of the 70 grams/m^2 of foliage that dies annually plus 5 gm/m^2 of dead roots remain as the annual increment of soil organic matter, for a total of 34 gms/m^2. The annual input of soil organic matter is also subject to further decomposition at a much slower rate. Studies by Jenkinson (1965) at soil temperatures of 24^o C indicate that the peat and humus may have a half life exceeding 24 years. Meadow soil temperatures are typically less than 16^o C, and half life must be much longer.

The chronology of soil accumulation in meadows provides information on the rate of organic soil accretion, which can be used as a check against the theoretical input of soil organic matter estimated above. Meadows typically build up 3 cm of soil per century. The soil is commonly 5 per cent organic by weight. Assuming a density of 1 gm/cm^3 for the soil organic matter, the soil build-up implies an annual increment of stable organic matter of:

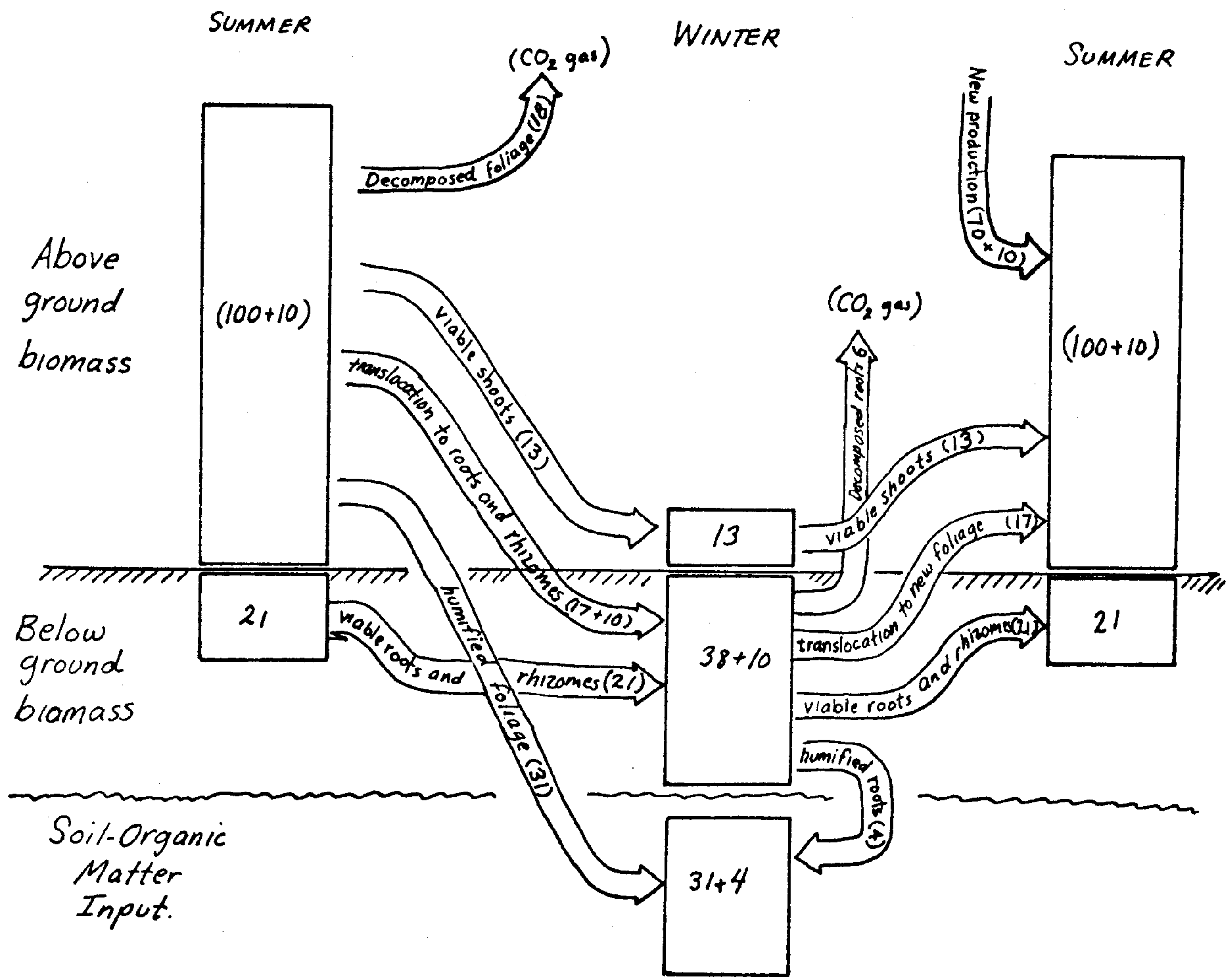


Figure 2-5. A theoretical estimate of the organic-matter budget of a Sierra meadow based on data from a sedge wetland in Minnesota (Bernard, 1974) scaled down by a factor of 100/852 to the 100 gm/m^2 above-ground biomass of a Sierra meadow.

$$(3\ \text{cm}^3/\text{cm}^2) \times (1\ \text{gm}/\text{cm}^3) \times (0.05) \times (1/100\ \text{yrs}) \times (10^4 \frac{\text{cm}^2}{\text{m}^2})$$
$$= 15\ \text{gm}/\text{m}^2\ \text{year}$$

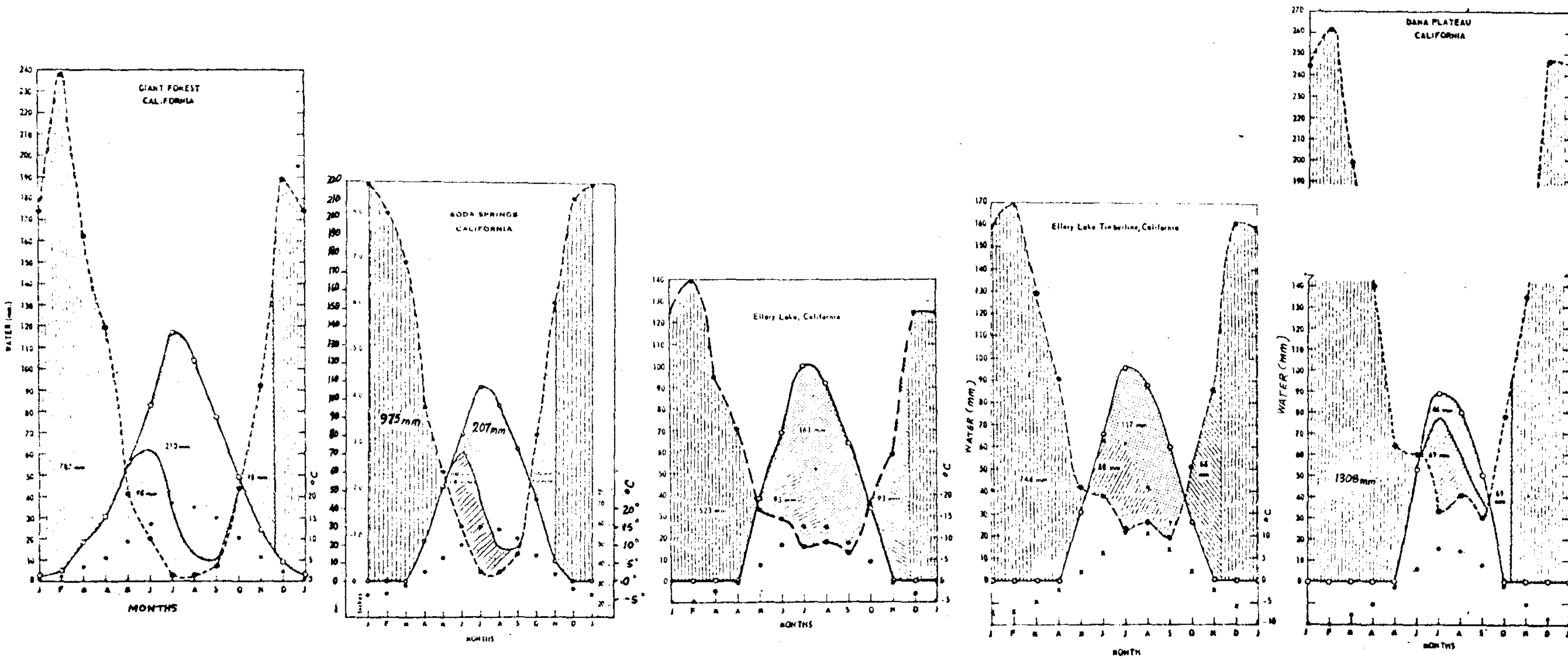
This figure compares reasonably with the estimated 34 gm/m^2 of annual peat and humus input, if we allow half of it to undergo long-term decay.

Seasonal climate and hydrology

Great seasonal contrast in climate distinguishes the Sierra Nevada from many mountain ranges of the world. Winter in the montane and subalpine belts is characterized by an 8-foot snowpack. Snow begins to accumulate in late November, and disappears in the latter part of May (figure 2-7). In contrast, the summers are clear, warm and dry. Summer rains are few and erratic. Total summer precipitation (June through August) averages less than 1 inch at sites below subalpine belt. When autumn arrives, the night temperatures often drop below freezing. Total autumn precipitation is light, averaging 6 inches, occurring largely as rain or snow that melts soon after falling. In December and January, snowpack accumulates rapidly (figure 2-7). Snowmelt is rapid during May, and the average date when montane basins are snow free is May 20th. These climatic data are summarized in figure 2-6: data from Giant Forest are typical of the southern Sierra montane belt. Higher subalpine areas (Ellery Lake, figure 2-6) are subject to colder harsher winters and considerably more summer rain which averages 2.4 inches delivered by thunder showers along the crest of the range.

The great seasonal contrasts in climate are reflected in groundwater and surface-water hydrologic conditions at meadow sites. Collection of detailed hydro-meteorological data by the Central Sierra Snow Laboratory for the period December, 1948 through December, 1950 (U.S. Army Corps of Engineers, 1951, 1952, and 1953) aids greatly in understanding the hydrologic processes on meadows. Precipitation, air temperature, snowpack, stream runoff, and ground-water levels were monitored continuously for two years on and around a 6-acre, wet, sedge meadow at the lower end of a willow-covered flat in a subalpine valley at 7350 ft. elevation (figure 2-7).

On the valley side, above the meadow, the water-table record (figure 2-7a) shows that the first major precipitation in November begins ground-water recharge of surficial soils and shallow bedrock joint systems. Early stages of snowmelt in May bring the valley sides to saturation. Excess water then runs off rapidly and seeps to the valley bottom



LEGEND

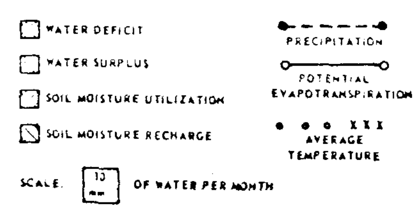
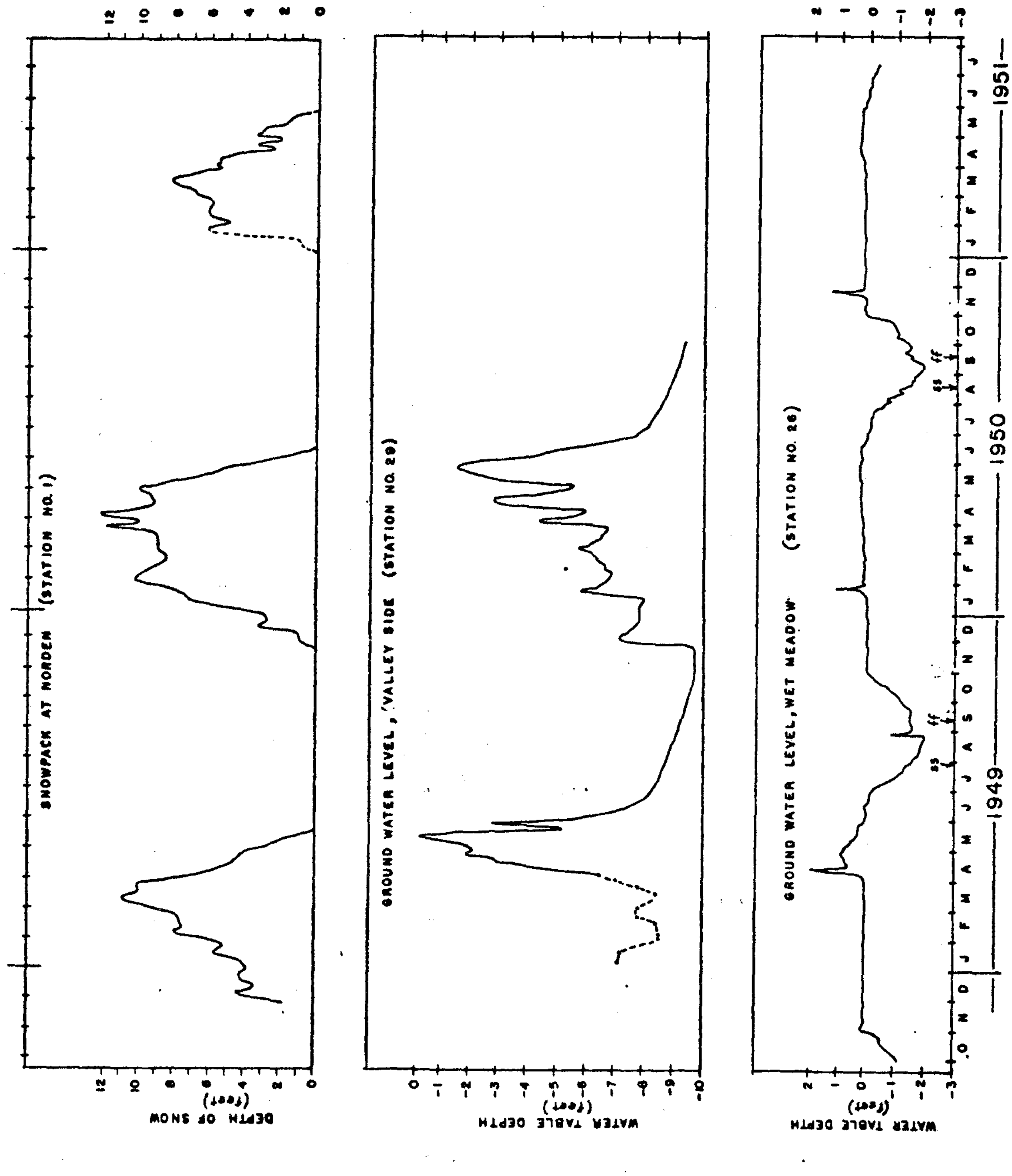


Figure 2-6. Climatic diagrams for the upper Sierra (after Major, 1965)

Figure 2-7. Seasonal changes in snow depth and ground water levels at the Central Sierra Snow Laboratory. Station No. 26 is a 6-acre meadow. Station No. 29 is a grus-mantled slope with scattered conifer trees. Station No. 1 is a clearing in the forest by the village of Norden, California. Date when stream runoff has stopped is indicated by "ss". Date of first autumn freeze is indicated by "ff". Spikes on the meadow water-table record are caused by filling of the well or ponding during rainstorms delivering over one inch of precipitation in a two-day period. (Data from Hydrometeorological Log of the Central Sierra Snow Laboratory, 1948-49, 49-50, 50-51 water years, U.S. Army Corps of Engineers, San Francisco, 1951 and 1952).


SNOWPACK AT NORDEN (STATION NO. 1)
DEPTH OF SNOW (feet)
GROUND WATER LEVEL, VALLEY SIDE (STATION NO. 29)
WATER TABLE DEPTH (feet)
GROUND WATER LEVEL, WET MEADOW (STATION NO. 26)
WATER TABLE DEPTH (feet)
1949
1950
1951

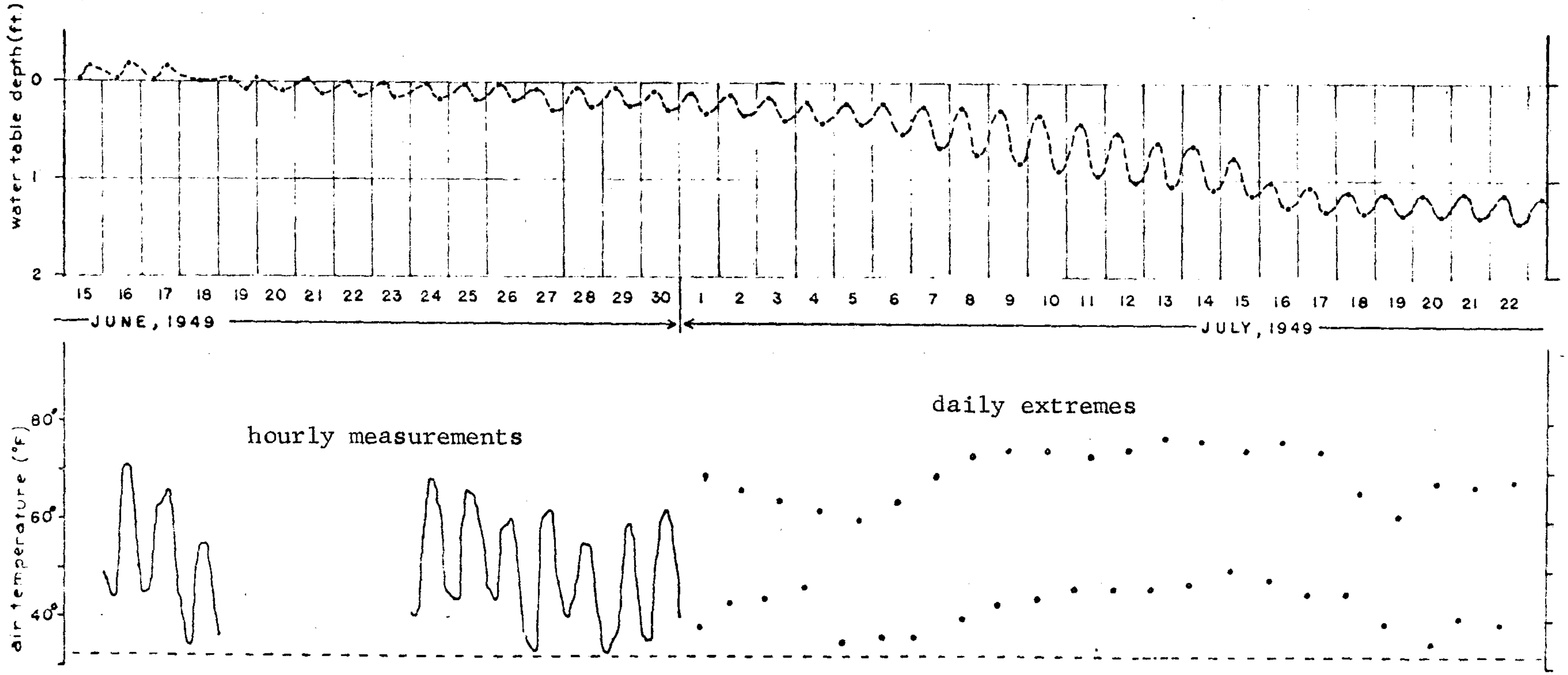


Figure 2-8. Daily ground water level and air temperature at a meadow near the Central Sierra Snow Laboratory (elev. 7300 ft), Yuba River Basin, California. Water levels above the surface are runoff from snowmelt. After snow has melted from the basin, in late June, the water table declines, and the large diurnal fluctuations in water level are caused by daytime evapotranspiration of the meadow plant community. (Data from the Hydrometeorological Log of the Central Sierra Snow Lab., 1948 - 49 water year, U.S. Army Corps of Engineers, San Franscisco, California, 1951)

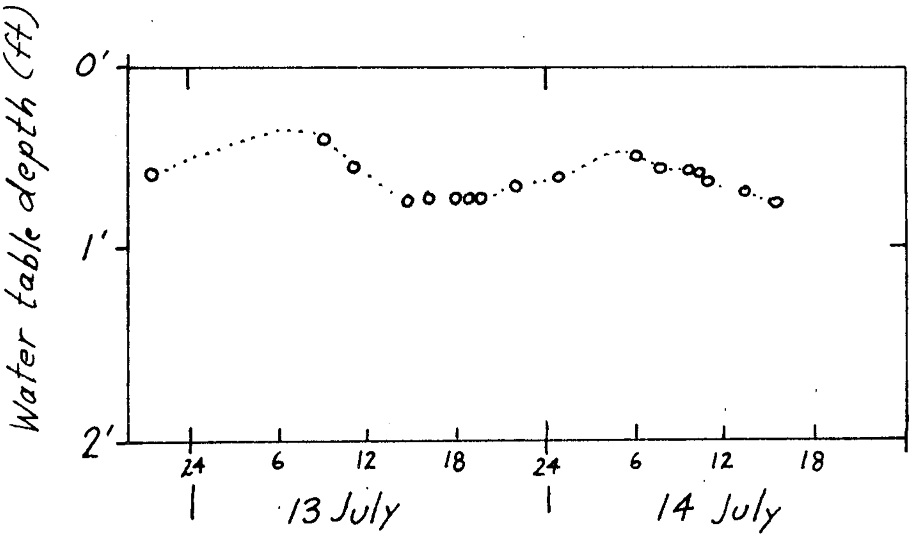


Figure 2-9. Diurnal fluctuation of water-table depth at weather station, Tuolumne Meadows, Yosemite National Park, California, July, 1974.

until the end of June, helping to sustain runoff in the streams. The entire basin is nearly snowfree by early July, and ground-water levels on the basin sides undergo a steady decline until November when late autumn rains begin to recharge the slopes. The summer decline in water-table level is attributed largely to gravity drainage of the slopes as water seeps off to the meadow below. Transpiration losses by scattered conifer thickets or riparian vegetation on the valleyside are apparently minor compared to downslope drainage, for no diurnal fluctuation timed with daytime transpiration appear in hourly water-table records.

Seepage and runoff from the slopes converge onto meadow sites in the valley bottom. The water table behaves differently in the meadow than on the slopes. During the period of snow cover and snowmelt, meadow soils are water saturated, and the water table is at the surface until the end of June (figure 2-7). During the latter stages of snowmelt, diurnal fluctuations are dominated by an afternoon maximum height near the surface caused by daytime snowmelt greatly exceeding the evapotranspiration (figure 2-8). The stream channels are full. By early July, direct runoff and seepage from melting snow has ceased, and the meadow area is being supplied by the drainage of ground water from the side slopes. The meadow is now totally dependent upon ground water. Fine silt topsoils draw water by capillarity to within reach of plant roots. From early July until early September, in contrast to the valley sides, the water table undergoes rapid drawdown, caused by the heavy evapotranspiration of the plant community. Vigorous plant transpiration during the day lowers the water table 0.2 to 0.55 feet to a minimum water-table height each afternoon. During the night, the water table is recharged to 80 or 90 per cent its previous maximum height which it reaches shortly before sunrise (figures 2-8 and 2-9). The water table declines in this cyclic manner through July and early August at a maximum rate of 0.085 feet per day.

Stream flow through the meadow ceases in the middle of August. Thereafter, the stream is dry except during occasional storms. By the end of August, the water table beneath the meadow has declined to a depth of two feet (figure 2-7a). In early September, the water table begins to rise at a rate of about 0.04 feet per day even though no significant precipitation falls until the end of October. By the end of October it is at the surface again. Since no corresponding reversal has occurred upon the slopes, the cause must lie in the meadow. Since the reversal occurs in advance of the first freeze, browning the foliage by frost to stop transpiration has not occurred. The foliage condition of the particular meadow in which ground-water was taken is not stated; however, meadows commonly tend to brown prior to the first freeze. Thus, the most likely explanation is that reduced transpiration is brought on by senescence of the plant community at the close of summer.

In conclusion, it is clear that the summer water table decline in meadows is caused almost entirely by the heavy evapotranspiration draft of the plant community, and not by drainage from the meadow. A sufficient inventory of water still exists upon the slopes at the end of summer to restore the meadow water table when the evapotranspiration demand decreases through browning of the vegetation. This browning is attributed primarily to physiological processes related to seasonal changes such as shorter days and cooler temperatures,and it is, of course, accelerated by the first episode of freezing.

Evapotranspiration in meadows

The relatively large diurnal fluctuations in meadow water-table levels during July indicate a considerable evapotranspiration (EV) rate from the meadow. These diurnal fluctuations can be evaluated to give estimates of EV loss rates and seepage recharge rates in the meadow. The following equation modified from Kittredge (1948) expresses the relation between the daytime water-table draw down and the rates of EV and seepage recharge.

$$(q_t - q_r)t_s = S_y h_l$$

Where:
q_t = hourly EV rate (ml of water/cm^2 hr)
q_r = hourly recharge rate (ml of water/cm^{2}hr)
t_s = hours of sunlight, which is considered to be 12 hours for sites in the Sierra June through August
S_y = specific yield of soil in which the fluctuations occur. In this study, it is the measured volume of water that drains by gravity from a water saturated soil divided by the volume of the saturated soil.
h_l = daytime water-table drop

The night time EV rate is assumed to be negligible, so that the equation for the height of water table recovery depends only upon the seepage flux.

$$q_r \ t_d = S_y \ h_2$$

where h_2 = night time water-table rise

t_d = hours of darkness

The seepage rate is assumed to be constant over the 24 hour period, although it may vary slightly because of the difference in head provided by the larger water-table fluctuations. The computed rates depend critically upon the value of the short-term specific yield. Specific yields were not measured for the soil column in which the water-table fluctuations are observed. Volume of water drained overnight from a typical sandy-loam meadow soil gave a specific yield (Sy) between 4 and 5 per cent, although Sy values in meadow soils could probably range from 2 to 10 per cent, an important uncertainty. Computation of the data in figure 2- for normal days when the maximum afternoon temperatures are between 60° and 70° F give EV rates, on a daily basis, of 0.5 to 0.7 ml/cm^2 day. During an exceptionally hot week when temperatures exceeded 74° F, the computed rate reached 1.3 cm/cm^2 day. Computation of data from subalpine Tuolumne Meadows (figure 2-9) yields a rate of 1.1 ml/cm^2 day. For a 90-day growing season, assuming an average rate of 0.6 ml/cm^2 day, the water comsumption for a montane meadow is 21 inches. This is nearly twice the 90-day potential evapotranspiration computed by Major (1965, shown in figure 2-6) from meteorological data at montane and subalpine elevations.

The average July recharge rates vary between 0.4 and 0.6 ml/cm^2 day but may reach 1.1 ml/cm^2 day, depending on the head provided by the water-table drop of the previous day. During September when EV is slight, the recharge rate is 0.4 ml/cm^2 day.

Two studies have measured EV from alpine meadows, but there is no comparable data from the mid-elevation sites. Mooney and others (1965) report EV losses from sedge-sod blocks over a 3-day period in July when daily temperature maximums were 70° F. The measured 0.3 ml/cm^2 day is comparable with the potential EV computed by Major (1965) for the Ellery Lake timberline site (figure 2-6). Klikoff (1965) measured EV rates throughout the summer on sod blocks from a timberline wet-meadow community near Ellery Lake and obtained much higher values, 0.6 to 0.8 ml/cm^2 day, comparable with the values obtained in this study.

Potential EV is defined as the water lost to the atmosphere from a low, complete plant cover which is actively growing and fully supplied with water (soil-water near field capacity). These conditions exist in wet meadows, yet the actual EV appears to be at least twice as large as the computed potential EV. Part of the discrepancy arises from the Thornthwaite (1948) method used to compute the potential EV which employs the mean monthly temperature, yet in mountain climates the daytime celsius temperature may be twice the night-time temperature. Averaging the temperature will cause an underestimation of at least 10 per cent. Still, this does not explain the large observed discrepancy which must be attributed to the inappropriate empirical constants for meadow sites in the Thornthwaite formula.

Surface-water hydrology and fluvial processes

During heavy rains or snowmelt, streams carry runoff water onto the meadow. This surface water may flow across the meadow either in a mode of retarded overland flow through the meadow foliage, or it may collect in well defined channels. Two parameters, drainage basin area and meadow slope, are important in determining whether or not a stream channel is established and maintained across a meadow. Meadows with drainage basins smaller than 0.8 mi^2 and surface slopes less than 2 per cent do not commonly have a through flowing stream channel. The various tributary water courses simply emerge on the meadow and spread out over the surface in overland flow. In reaches of the meadow where the slope exceeds 2 per cent, channels form, but they are small and often vegetated with sedge. Upon succeeding to a reach with a shallower gradient, the channels end and the water flows overland again. Streams draining areas greater than 0.8 square miles commonly have well defined channels across meadows.

Channel pattern of streams in meadow vary from straight (sinuosity[3] = 1) to gently meandering (sinuosity = 2.1). The pattern seems to depend upon the drainage area and meadow gradient, but a detailed analysis has not been carried out. Small drainage areas and steeper meadow gradients tend to produce straight channels. Large catchments and gentler gradients tend to produce more sinuousity. Abandoned meander loops are observed where large streams flow through relatively flat meadows, and it is clear that such streams have shifted course a number of times.

Channels through meadows are rectangular in cross section being confined by steep banks of slightly cohesive sod-bound topsoils that extend to the mid-summer water line.

footnote for page 32

[3]Leopold and others (1964) put the meandering nature of channels on a quantitative basis. Sinuosity is defined as the ratio of the length of stream channel to the length of valley. Streams with sinuosity greater than 1.5 are considered meandering. Sinuosity reaches 4 or more for some rivers of the world.

Channels are 6 to 10 times wider than they are deep, and most have a gravel bed. This bed is locally armoured by rust-stained gravel 4 to 6 cm in diameter, and bars of finer gravel and transported grus are common. The bed configuration produces a water surface characterized by continuous riffles along the steeper armoured reaches,and riffles and pools in the reaches with gentle gradients. At the outside of bends and around large rocks, pools are scoured to depths of 2 to 6 feet.

Channels appear to be quite stable. Jahnda (1967) concludes that transport of coarse gravel and channel changes do not take place annually but occur during major floods with recurrence intervals greater than 10 years. Large streams with migrating meanders undercut topsoil banks at outer bends of meanders annually, but deposit gravel and sand at inner bends only during infrequent major floods. Meadow vegetation slowly invades the gravel bars.

Downcutting of channels through meadows is usually limited to a few feet. Gravel layers within meadow deposits act as local base levels of topsoil erosion.

The considerable thickness and continuity of topsoils and alluvium beneath montane meadows attests that these sites are aggrading. Erosion creates only minor modifications of the depositional surface.

Two major types of clastic material are deposited on the meadow surface. Silt to medium-sized sand grains are transported by overland flow through the sprouting sedge blades during annual spring snowmelt flooding and occasional summer thunder showers. This leaves thin discontinuous silt and sand lenses on the meadow surface. Wind transported dust is caught by the foliage during the summer. In the fall the dead foliage, including the dust, is incorporated into the meadow topsoil.

The second type of clastic material deposited upon meadow surfaces comprises sheets of sand and gravel composed largely of transported grus . Some is delivered from side tributaries, but most is brought onto the meadow by the main stream and it is spread over the surface during overbank flooding. As the stream discharges from a steeper confined reach in bedrock, it appears to transport its bedload as a transverse bar which migrates across the meadow if flood flow is sustained. These bars are usually 0.5 to 1 foot thick. Gravel bars commonly seen on meadows in the southern Sierra during the 1970's were probably deposited by the December, 1966 flood (figure 3-13).

Another type of clastic deposit one might anticipate within meadow stratigraphic sections would be abandoned, armoured stream courses left as the stream meanders. This type of deposit is rare on surface of meadows and in the underlying stratigraphy, suggesting that the channels are stable.

In order to interpret the surficial clastic deposits we must examine the frequency and nature of major floods in the watersheds and the resulting hydraulic regime of flood flows across meadows.

Floods in the Sierra

Major floods from snow-covered mountains often occur with the spring snowmelt, but the Sierra is unusual in that its greatest floods have occurred in the dead of winter "when one would suppose all the wild waters would be muffled and chained in frost and snow" (Muir, 1894). These great floods are rare occurrences and none has happened since 1966. They are caused by intense rainfall from unusual frontal-type storms of regional extent in December, January or early February, and are attributed to warm, moist, air masses from the tropical Pacific. Along a stretch of the Sierra west slope 100 to several hundred miles long, over 20 inches of rain falls in a few days, and in local areas, 50 to 100 miles long, up to 30 inches may fall (figure 2-10). Intense rainfall is limited to a particular stretch of range, thus extreme flooding in the northern Sierra may not be accompanied by a similar event in the southern Sierra. Torrential rains melt pre-existing snow cover and cause flooding in valleys of the foothills and montane belts. Above 8500 feet precipitation falls largely as snow. Presumably these floods are responsible for deposition of sheets of sand and gravel on the montane meadows.

How often have meadows been subjected to such great floods? A traditional answer can be obtained by computing the recurrence interval (T) of events ranked in order of magnitude (m) from a record of length n-years. The procedure most commonly used is given by Dalrymple (1960).

$$T = \frac{n + 1}{m}$$

A time scale (y) has been devised based upon extreme value theory that tends to make the frequency curve plot a straight line. The equation for the time graduation is:

$$y = -\ln\left(-\ln\left(1 - \frac{1}{T}\right)\right)$$

For comparison purposes, the flood magnitude may be defined as the peak discharge divided

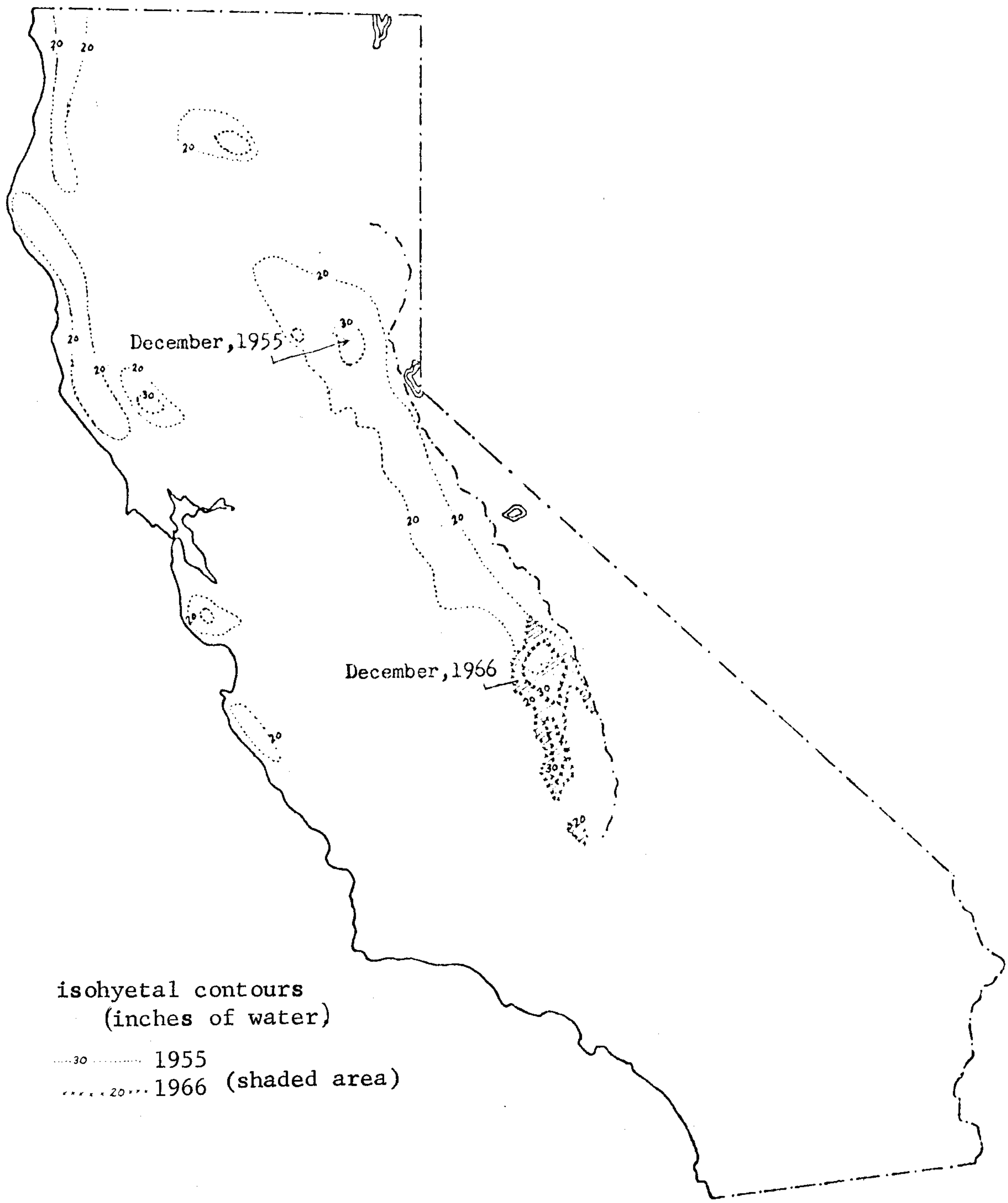


Figure 2-10. Comparison of isohyetal maps (total rainfall) for the December 17 - 26, 1955 storm with the December 2 - 6, 1966 storm showing that different storm tracks of warm moist Pacific air masses led to torrential rain distribution in very different parts of the Sierra (data from California Division of Water Resources, 1956; Dean and Scott, 1971; and U. S. Geological Survey, 1967).

by the upstream drainage area. Flood frequency curves are fitted to the plot of magnitude versus the recurrence interval. The recurrence interval read from the curve is a statement of probability. Thus a recurrence interval of 10 years applies to an event that has a 10 per cent chance of recurring in a single year (Leopold and others, 1964). The method gives useful results within the period of record but it cannot be safely extrapolated.

Jahnda (1966) analyzed flood flows from several Sierra streams with long stream-gage records. He identified two distributions of annual flood events (figure 2-11). The first distribution contains the annual floods, usually of low magnitude, on all of the streams and also the rare floods of the high subalpine streams. These events are temporaly related to spring and early summer snowmelt.

The other distribution contains the great winter floods on streams in the montane and foothill belt. Their greater magnitude produces a curve that sharply diverges from that of the snowmelt floods. These are not frequent events, and Jahnda (1966) concludes that their recurrence is measured in tens of years.

Young and Cruff (1967) analyzed records of 90 gaging stations on the west slope of the Sierra and incorporated historical information on the magnitude of great floods in order to determine the flood frequency curves for 10, 25, and 50-year events. By multiple regression analysis they determine equations relating the maximum flood discharge in cfs/mi^2 for the n-year event (Q_n) to the drainage area in mi^2 (A), the altitude index[4] of the stream in 10^{-3} feet (E), and the mean annual precipitation for the basin in inches per year (P). For these flood events they give the equations:

$$Q_{10} = 0.336\ A^{0.84}\ E^{-0.52}\ P^{1.74}$$

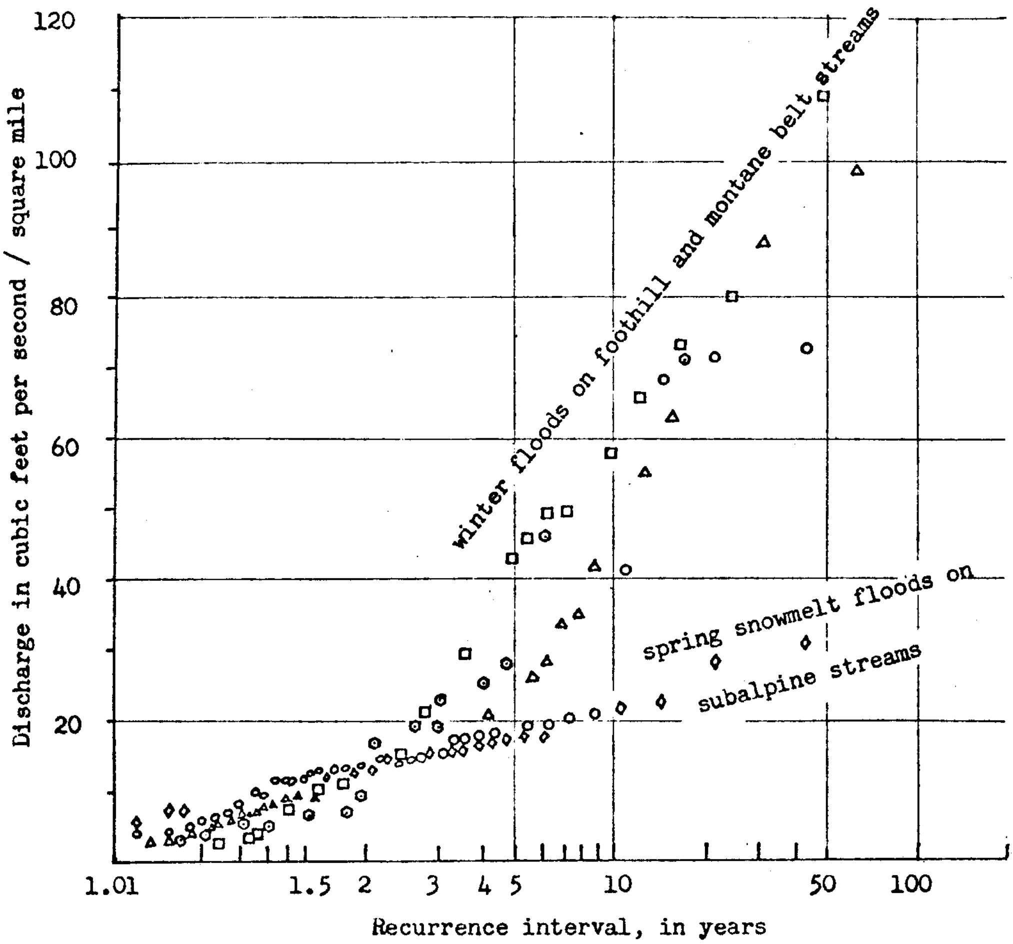
$$Q_{25} = 0.589\ A^{0.84}\ E^{-0.43}\ P^{1.67}$$

$$Q_{50} = 1.410\ A^{0.83}\ E^{-0.30}\ P^{1.49}$$

Young and Cruff (1967) caution that these curves do not apply to basins smaller than 10 square miles. The curves are plotted in figure 2-12 for a 10-square-mile basin and for a range of elevations and mean annual precipitation typical of the principal meadow belt. The curves are probably unreliable for higher elevations, as there are no winter gaging stations above 8500 feet. It was observed during the great storms of 1955 and 1966 that a continuous snow cover remained above 8500 feet. Rain falling above this elevation is absorbed by snow to produce a dense, iced layer with no significant water loss (Dean and Scott, 1971, and California Division of Water Resources, 1956).

footnote for page 36

[4]Altitude index is a rough measurement of the mean altitude of the basin computed by averaging the altitudes of the two points 15 and 85 per cent of the total channel length upstream from the gaging station (Young and Cruff, 1967).



		Drainage area	Altitude index
o	Merced River at Pohono Bridge (Yosemite)	321 mi^2	6440 ft
▲	Merced River at Bagby	912 mi^2	3980 ft
□	Fine Gold Creek near Friant	93 mi^2	1560 ft
◊	Bear Creek near Thomas Edison Dam	54 mi^2	9340 ft
●	Sagehen Creek, Truckee River drainage	11 mi^2	7040 ft

Figure 2-11. Flood frequency curves for selected stations in the central Sierra Nevada, California (after Jahnda, 1966, and A. Cunningham, unpublished data)

The storm of December 5-6, 1966 produced a great flood in the southern Sierra. On the Tule River, the discharge was 3.5 times the probable 50-year flood (Dean and Scott, 1971). Large flood magnitudes were measured in drainage basins as small as 1.2 mi^2 and as large as 102 mi^2. On the Kern and Kaweah, the maximum stages may have been just short of the greatest historical flood of the southern Sierra that occurred in December of 1867. These large peak flows, as well as the largest flow of the December, 1955 flood, are plotted on the flood frequency graph (fig. 2-12)containing the flood frequency curves derived by Young and Cruff (1967). The historical record of the Sierra probably extends back to 1850, therefore these flows are plotted along a 125-year recurrence line, since it is not known what their true recurrence interval should be.

The greatest recorded flood magnitude (575 cfs/mi^2) from the west slope of the Sierra occurred in the December, 1964 storm on the south Honcut Creek in the Feather River Basin draining an area of 30.6 square miles with an altitude index of 1260 feet. This flood in the northern Sierra may have been exceeded by that of January, 1862. Discharges above 300 cfs/mi^2 are somewhat more common in northern California streams, but nevertheless, they are rare events on any stream.

Such great floods have a recurrence interval between 50 and 200 years. Is it possible that still greater floods have occurred in the past and can occur in the future? Hoyt and Langbein (1955) have reviewed the greatest flood discharges recorded in the United States. Flood discharges an order of magnitude larger than those of the Sierra have occurred elsewhere associated with regional storms, notably in the southern Appalachians and in south central Texas. A local spring cloudburst on the east side of the Cascade Range, near Waterville, Washington delivered 30,000 cfs from a 5.2 mi^2 watershed. Floods of similar magnitude are known from convective thundershowers in the California deserts. Yet no such floods (6000 cfs/mi^2) are known from the Sierras, and their probability is conjectural.

Floods between 300 and 600 cfs/mi^2 have occurred in the Sierra, and they have a probability of recurring at least every few hundred years. The hydraulic nature of a 600 cfs/mi^2 flood flow over a montane meadow can be estimated using the Manning equation for flow in an open, rough, channel (Appendix C). For a meadow 200 feet wide across the valley bottom, a gradient of 0.02, an upstream drainage of 3 mi^2 and a channel roughness of 0.04, the depth of flow for such a flood would be 1 foot. The

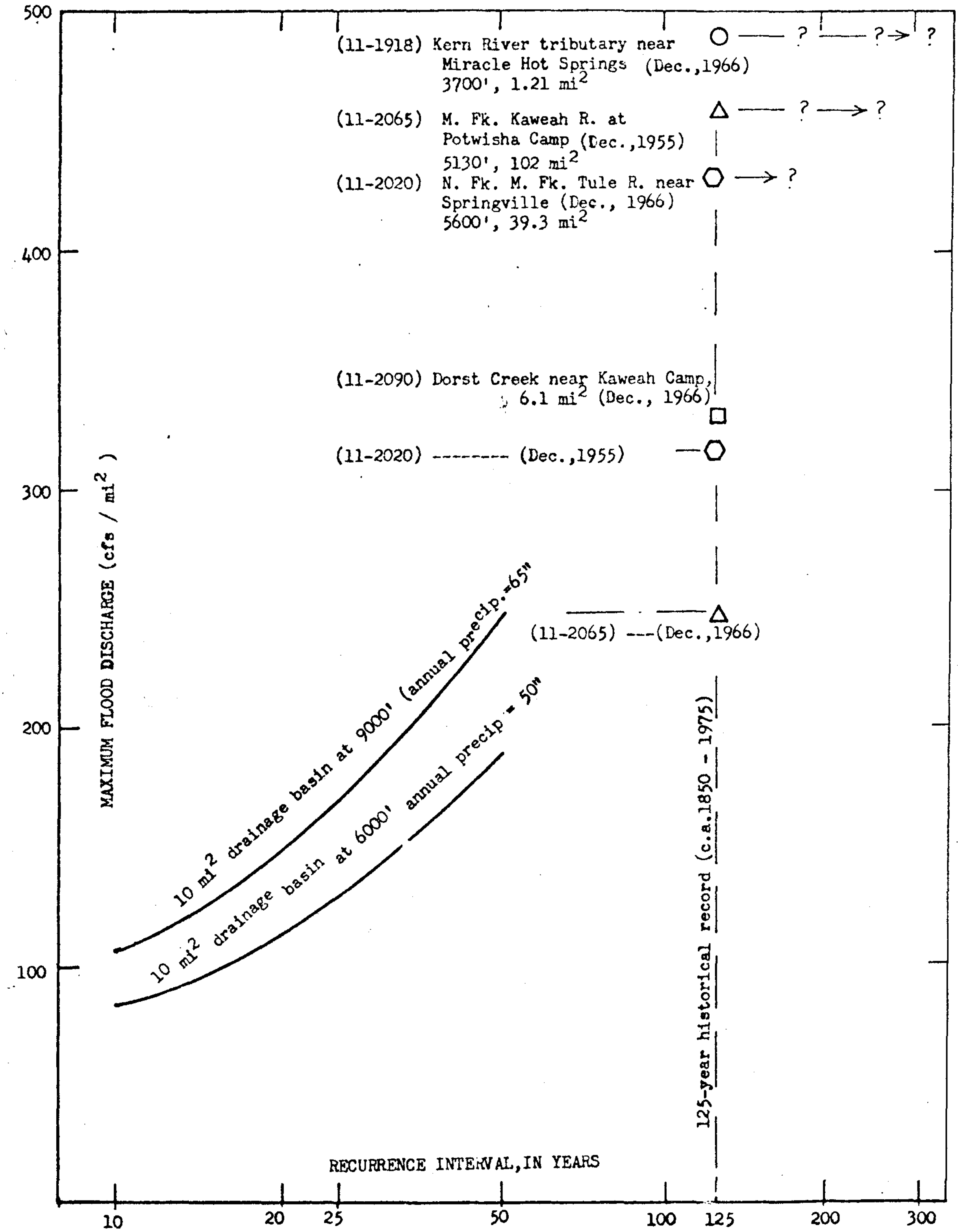


Figure 2-12. Flood frequency plot for small drainage basins on the west-slope Sierra showing curves derived by Young and Cruff (1967) and data on major floods from Dean and Scott (1971).

mean flow velocity would be 6 feet per second. The competency, or the size of the largest particle that can be transported, is computed from Shields criteria (Blatt and others, 1972, and Vanoni, 1964) and the Duboys equation (Chow, 1959) for the traction on the bed. This flood flow is capable of transporting a 5 cm diameter particle along the slope of the meadow. If the flow is to some extent channelized, a larger particle might be transported. The maximum likely flood (6000 cfs/mi^2) would flow 5 feet deep over the meadow at a mean velocity of 18 feet per second. This flow is competent to initiate the motion of boulders. No boulders are found on the meadows or in their Holocene deposits indicating that these floods have not occurred in the last few thousand years, on the west slope montane belt of the Sierra.

A recently constructed gravel bar on a meadow in Kings Canyon National Park is shown in figure 2-13. The 2-mi^2 drainage above the meadow almost certainly produced flood flows across part of the meadow during the December, 1966 storm. The bar is composed largely of transported gruss with clasts up to 2 cm. This size of clast implies a bed traction of 200 dynes/cm^2 (Blatt and others, 1972). By means of Duboy's formula the required depth of flow on a 0.02-gradient slope is 0.3 feet. From Manning's formula the discharge across the one-hundred foot width of the meadow would be 431 cfs, or 215 cfs per square mile. The graph on figure 2-12, suggests that this was about a 50-year event.

Figure 2-13. Gravel bar on an un-named meadow, 1.2 miles northwest of Seville Lake, elevation 8300 feet, Kings Canyon National Park, California. Photo taken on July 19, 1973.

CHAPTER III

THE STRATIGRAPHIC RECORD

Introduction

A similar succession of topsoil and alluvial layers is exposed beneath meadows by Twentieth-Century dissection at a number of sites within the principal meadow belt (figure 1-1) on the west slope of the Sierra Nevada. Detailed study of well-exposed sections reveals strong correspondence in the character and sequence of these layers. In basal deposits of five, widely spaced, dissected meadows, is a distinctive paleosol. The similarity of the profiles in this buried soil and close agreement of radiocarbon ages from the accreted "A" horizon at all five sites indicate a short (1500 years), but effective, soil forming episode. Deposits in the middle part of the meadow sections display greater local differences in sediment character. They are interpreted principally as deposits of dry meadows, sparsely vegetated gravel flats, or coniferous forests and as slope-wash alluvium. In all five sections the uppermost 2 to 4 feet contain peat and humic topsoil layers interstratified with sorted sandy gravel. At two of the sites, the upper section includes an abrupt transition from a forest environment to an open meadow.

The following descriptions of site settings and stratigraphic relationships combined with observations of processes occurring in modern meadows (discussed in chapter 2) provide a basis for conclusions concerning the origins of these particular meadows, and the nature of processes that maintain mountain meadows. Information is also obtained on the timing and intensity of Holocene climatic variations in the Sierra Nevada.

The best sections are exposed in gully walls within East Meadow, Beasore Meadow, Cabin and Exchequer Meadows, and in Boggy Meadow, locations of which are shown in the index map (figure 1-1). Detailed locations are tabulated in Appendix A. A discussion is included on meadows in the extensively glaciated sub-alpine belt of the range, based largely on investigations in the Tuolumne Meadows area. Deep dissection has not occurred in sub-alpine meadows, so the stratigraphic investigations are limited to shallow excavations, core and auger holes.

Setting and Description of the Sections

East Meadow, Yosemite National Park

Setting

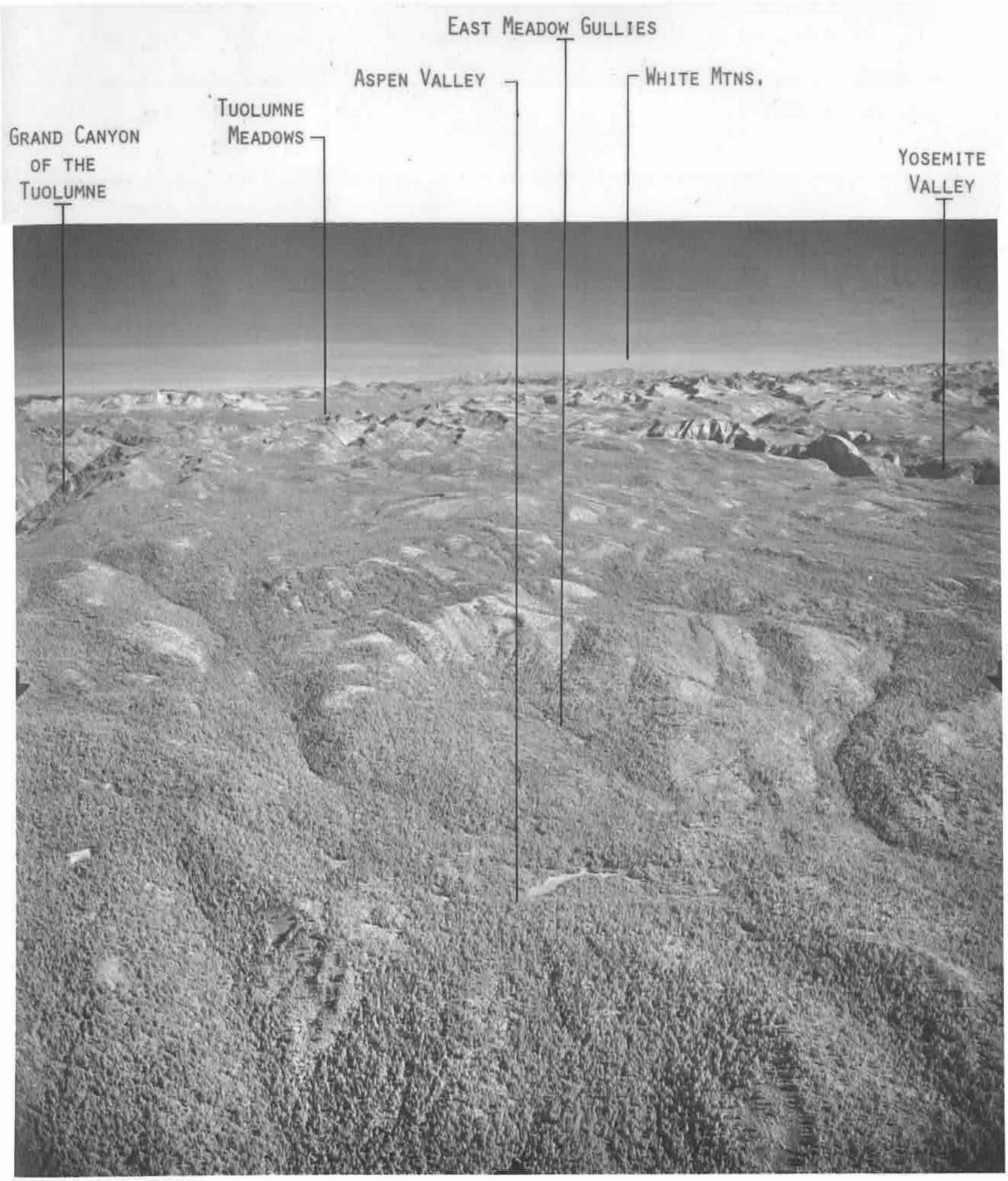
The most representative and best exposed stratigraphic sequence beneath a Sierra meadow is in the walls of a gully dissecting, and consequently destroying a 15-acre meadow in western Yosemite National Park. This is "East Meadow" of Aspen Valley which lies in an unglaciated embayment of the winding step front that rises abruptly from the 6000 to 8000-foot level on the west-sloping upland between Tuolumne (Hetchy Hetchy) and Merced (Yosemite) canyons (figure 3-1). Bedrock within the 3.5 square-mile drainage basin above East Meadow is coarse-grained biotite-hornblende granodiorite with zones of mafic inclusions. A pervasive fabric to the drainages reflects a system of vertical bedrock joints oriented N 55^{o} E. The valley of East Meadow is aligned along this strike.

The main stream which feeds the meadow forms in the thickly-forested valley bottom at the base of the rocky step. It flows one mile through forest to the meadow, two-thirds of a mile in the gully dissecting the meadow, and then a mile through a thickly forested reach before it cascades over a bedrock knickpoint. The stream above the meadow goes dry in mid-summer, but seepage into its gully maintains a perennial flow that dwindles to about 0.5 cfs by late summer.

Soils on the rocky slopes of the basin above the meadow are discontinuous patches of shallow grus upon exfoliating bedrock. On lower slopes and around the meadow disintegrated granodiorite accumulates by slope wash and is admixed with tree litter to form a coarse sandy loam in which the mixed-conifer forest is rooted. On the valley floor within the meadow, alluvial and organic deposits were accumulating prior to dissection on a smooth surface that sloped parallel to the stream with a gradient of 2.0 to 3.6 per cent.

The dry rocky slopes above East Meadow are covered with patches of Canadian zone mountain chaparall (Storer and Usinger, 1965, p. 27), Jeffery Pine, and a few Sugar Pine. Lower slopes are thickly forested with White Fir, Jeffery Pine, Lodgepole Pine, Sugar Pine, and Incense Cedar. Red Fir and White Pine are rare. On the forested valley bottom White Fir is most abundant. A few aspens are found around the meadow opening. The

Figure 3-1. Oblique air photo looking east over west-central Yosemite National Park. The barren step front rises 1800 ft (550 m) behind the dissected 15-acre East Meadow of Aspen Valley. The main meadow of Aspen Valley is a typical montane meadow of 20 acres. (U. S. Geological Survey Photo GS-OAH-3-23, 11/7/55).



East Meadow Gullies
Aspen Valley
White Mtns.
Tuolumne Meadows
Grand Canyon of the Tuolumne
Yosemite Valley

several willow clumps along the stream course through the meadow may have been introduced to check erosion. The meadow community is dominated by sedge with forbs and grasses accounting for probably less than 35 per cent of the cover. By early summer, sedge stands knee-high in the two north branching undissected wet meadows. (See index map in figure 3-2.)

Recent history

It is probable, but undocumented, that early sheepherders used the Aspen Valley meadows during the 1860s. The meadows were homesteaded in 1880 by Jeremiah Hodgdon. The Tioga Wagon Road was constructed in 1882-83 through Aspen Valley to mines at Bennetsville on Tioga Pass. According to Hodgdon's grandson, Robert Bright of Knights Ferry, California, hay was cut annually on East Meadow until 1910. In the winter or early spring of 1911, Bright recalls that the meadow was deeply gullied by floodwater from unusually heavy rains. This rainstorm must have been the January 31, 1911 storm that caused above normal stages on the Merced, Tuolumne, Stansislaus, Calavares, Cosumnes and Pine rivers (Hoyt and Langbein, 1955, p. 358; and Young and Cruff, 1967, p. 135 and p. 121).

A gully now runs the length of the meadow. Two headwater branches (figure 3-2) are currently encroaching upon the forested area above the meadow, largely by ground-water sapping. The slumped material produced by sapping is probably removed during floods. No major floods have occurred in East Meadow since 1952 as evidenced by an undamaged log roadway built across the gully during a 4-year selective logging operation that ceased in 1952.

Proximity to Pleistocene glaciers

This area lies outside the limit of Wisconsin glaciation shown by Matthes on an unpublished map compiled for Yosemite National Park. Wahrhaftig and Birman (1965) locate the Wisconsin climatic firn line, from elevation of the lowest cirques, at 8000 feet in this region which is too high to generate a glacier in East Meadow. A thorough search of the valley has revealed no erratic rocks, till, or other features of glacial origin supporting the conclusion of no Wisconsin glacial activity.

If, during older and more extensive glacial stages, ice accumulated on the step front, or high-country ice spilled into the valley from the southeast, all evidence of such an invasion has since been removed. The El Portal stage ice shown by Matthes (1930) was more extensive than Wisconsin ice and its edge would have lain near, but not quite in, this valley.

The stratigraphic section

Presently, the main gully extends for 2500 feet through the meadow providing some exposure of the bedrock surface upon which the meadow deposits lie. This irregular surface has low spurs and knobs, and many core stones (Linton, 1955) 5 to 8 feet in diameter lie upon it. Most core stones are detached, but some, such as shown in figure 3-3, remain embedded in disintegrated granodiorite bedrock. Prior to development of the meadow fill, the stream apparently followed a course confined by spurs, knobs and core stones on a gradient of about 1 per cent. Near points "A" and "B" (figure 3-2) coarse, sub-angular gravels devoid of organic material, encountered by augering are thought to represent the bedload of this confined stream channel.

The basal depositional unit and paleosol

The initial organic deposits of the meadow sequence are interbedded muck and thick sand layers. The top of this unit can be traced laterally to a burial soil profile developed upon the bedrock knobs and spurs. Muck layers rest upon a three-foot thickness of inorganic fine sand under point "B" (figure 3-2) in the middle of the meadow. In the upper meadow the muck rests upon inorganic sandy gravels. The muck layers represent the first deposits clearly involving vegetation at this site. The muck is black, humic, fine, cohesive material containing abundant sticks and small logs. A few *in situ* stumps identified as cedar and lodgepole pine are rooted at or near its base, and at the top of the muck unit is a rooted fir stump.

Intercalcated sand layers are composed mostly of fine grained, silty sand, except near the upper end of the meadow (250 feet from point "A", figure 3-2) where gravelly sand lenses with clasts less than 1 cm are exposed.

At a level concordant with the top of the muck unit, but resting upon bedrock protrusions is the complete profile of a well-preserved buried paleosol. This paleosol abuts against core stones (figure 3-3) but its parent material is the granodiorite bedrock upon

Figure 3-2. Longitudinal section through East Meadow of Aspen Valley, western Yosemite National Park. 15 x vertical exaggeration.

LONGITUDINAL SECTION THROUGH EAST MEADOW
WESTERN YOSEMITE NATIONAL PARK, CALIF.

15 X VERTICAL EXAGGERATION

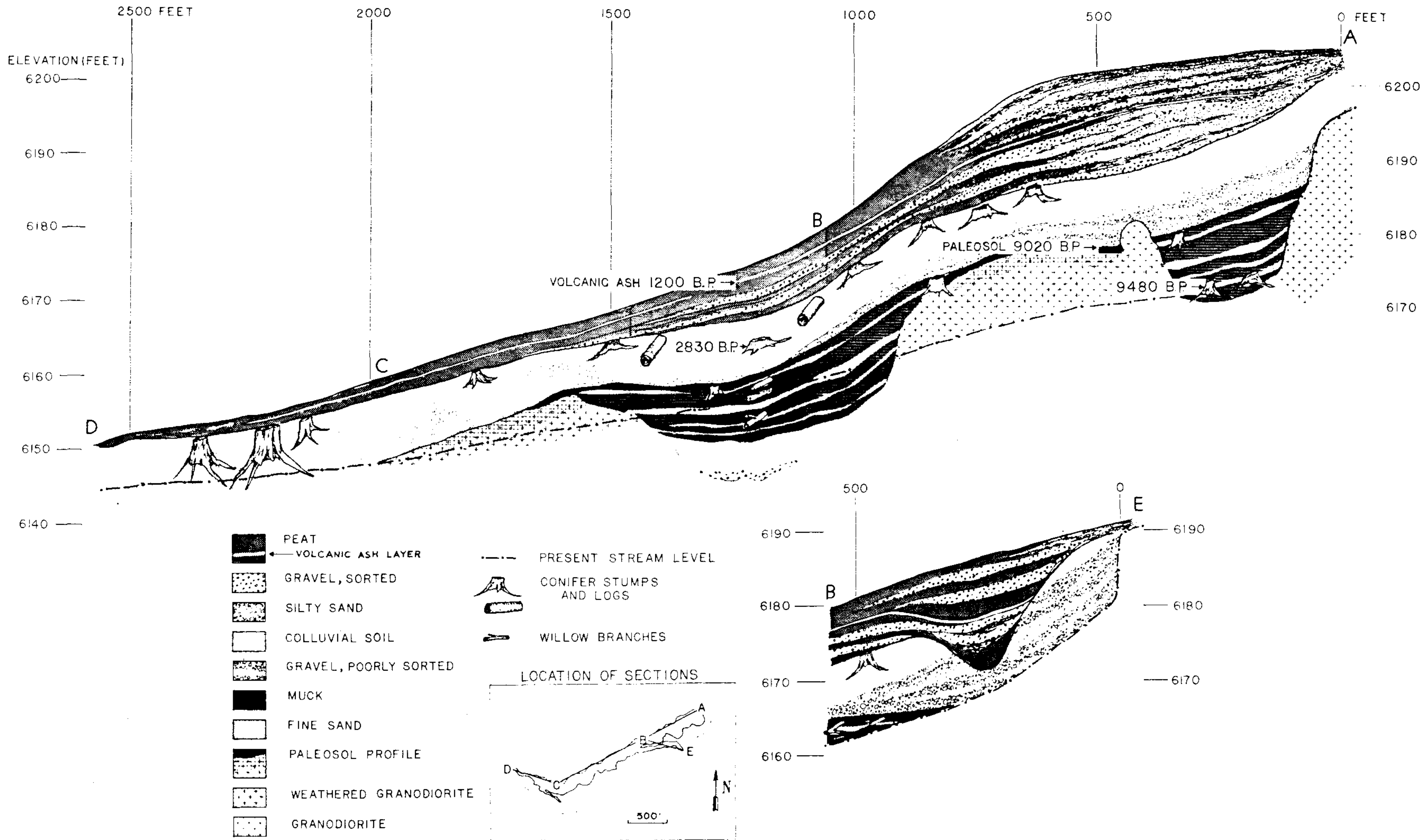


Figure 3 -3. Buried granodiorite core stones and the basal paleosol exposed in the gully of East Meadow, Yosemite National Park. The horizontal dark layer to the right of the 6-foot stadia rod and 1.5 feet below the top of the rod is the 9020 year B.P paleosol "O" horizon. The burrow of an introduced beaver colony (at the base of the rod) was dug into partially disintegrated granodiorite of the "C" horizon. A seven foot thickness of gravel and forest soils of the middle unit rests upon the core stone and paleosol. At and above the level marked by clumps of aspen saplings growing out from the bank are interbedded peaty meadow deposits and gravels of the upper unit which here attains a thickness of 17 feet.

Figure 3 - 4. The East Meadow stratigraphic section at point "B" on figure 3-2. The 12-foot stadia rod is extended for scale. The upper depositional unit comprises the top one-third of the section. Here it is composed of three, light colored, fine gravel beds separated by two peat beds. The lowest gravel bed pinches out locally to the right. A massive 3-foot peat layer caps the section and contains the 1200-year-B.P. volcanic ash layer which shows as a continuous thin white streak near the top of the peat bed.

Figure 3-3

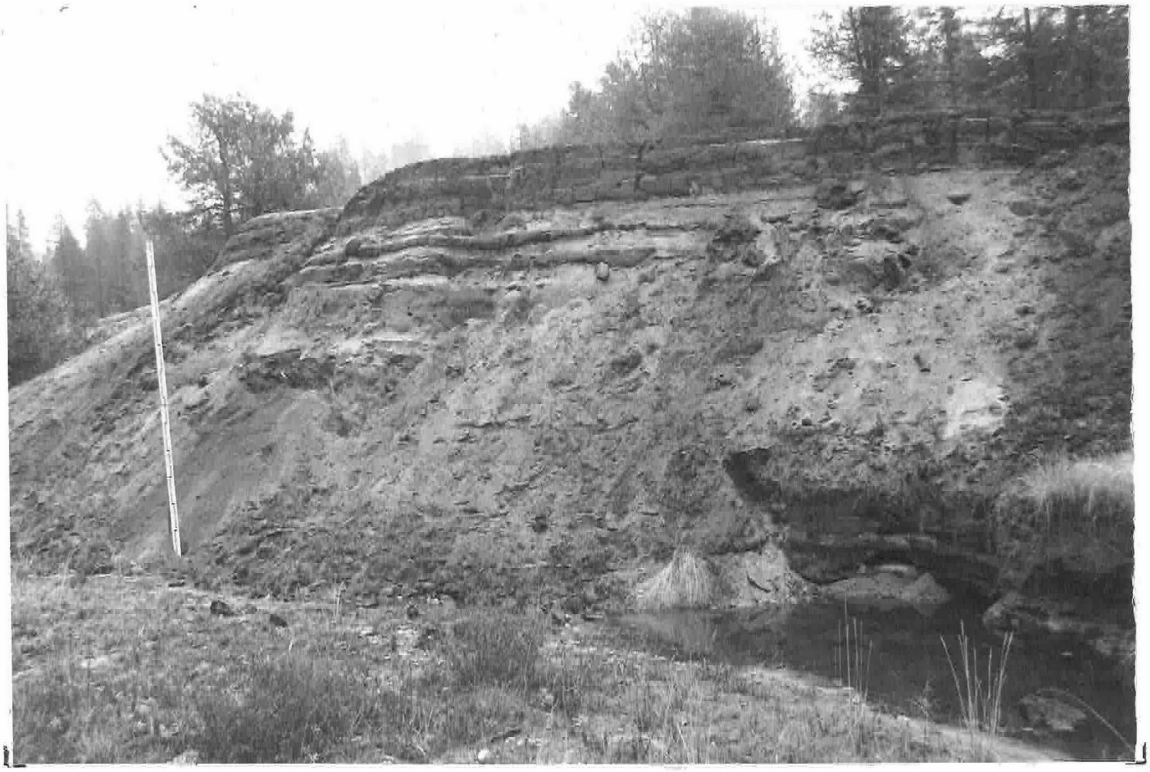
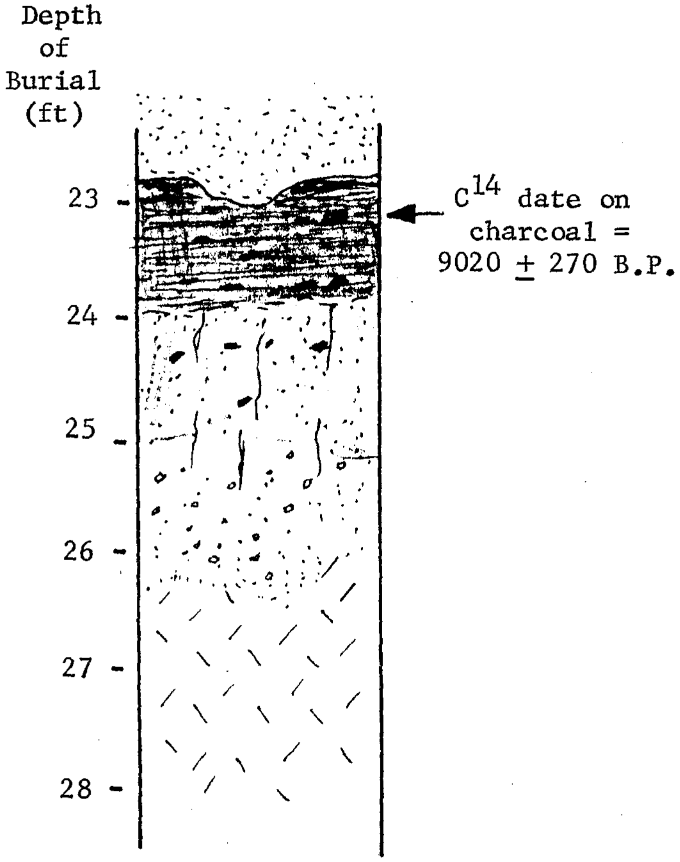


Figure 3-4



HORIZON DESCRIPTION

O: dark gray(10YR 3/1) finely divided humus and silt. Sticky but friable. Bark and charcoal abundant at top.

A: grayish-brown(2.5Y 5/2) sandy loam, friable, scattered charcoal throughout.

AC: pale olive(5Y 6/3) coarse sandy loam, slighty sticky but friable. Material is largely grus. Relict oxi-root channels penetrate this horizon

C: light gray(5Y 7/1) disinteg-grated granodiorite. Can be cut with spade. Primary igneous features are preserved.

Figure 3-5. Profile description of basal paleosol at East Meadow.

which it rests. The soil formed on the better drained rocky irregularities of the valley bottom while muck and fine clastics were accumulating in the intervening low spots and stream channels. A field description of the paleosol profile is presented in figure 3-5. Stumps identified as Jeffery Pine and fir are rooted in it (table 3-1) implying fair drainage for part of its history. The black humic "O" horizon is slightly browner than the muck layers, and contains considerably more charcoal. It is crudely stratified and is an accreted horizon presumably formed under moist, but moderately drained forested conditions. The underlying "A" and "AC" horizons contain a few rust stripes marking the sites of decomposed roots extending to a depth of 2.5 feet within the profile. At a downstream locality of the paleosol (station 1600 feet, figure 3-2) the "AC" horizon contains a few scattered sub-angular cobbles of aplite, apparently the remains of a thin dike that did not disintegrate with the granodiorite. These aplite cobbles have friable yellow-stained (10 YR 7/8) weathering rinds 3 to 5 mm thick which serve as a crude indication of the intensity of weathering undergone by a fine-grained felsic rock during this soil forming interval. Aplite cobbles in the overlying deposits have no visible weathering rind.

The "A" and "AC" horizons are differentiated primarily on the basis of color plus a slightly sticky consistency in the "AC" material. Admixed charcoal, and fine organic material give the parent gruss of the "A" horizon a slightly browner hue and darker color value than the "AC" horizon which has no visible organic component. There are 5-cm-diameter animal burrowings in the "A" horizon filled with "AC" horizon material.

Chronology of the basal unit: Two radiocarbon age determinations have been made on material from this basal unit. A charcoal layer, shown in figure 3-5 near the top of the "O" horizon of the paleosol yields an age of 9020 B.P. (UGa-447). A lodgepole pine stump rooted in gravelly sand beneath 9 feet of basal-unit muck and sand (figure 3-2) yields an age of 9480 years B.P. (UGa-452). These two ages confirm the field conclusion that the paleosol organic horizon and the muck layers accumulated in the same time interval.

The middle unit

Resting upon the paleosol or upon the uppermost muck layer are 8 to 10 feet of poorly sorted, rudely bedded, gravelly sand and loamy sand deposits. The lower contact of this unit is everywhere sharp, and it is commonly accentuated by a basal cobble layer with sub-rounded clasts up to 10-cm diameter resting upon the dark-brown muck or the organic horizon of the paleosol. The top of this middle unit is well defined by a line of truncated tree stumps rooted in a slightly cohesive coarse loamy sand material.

Layers in the upper half of this unit are coarse textured, consisting of transported gruss admixed with varying amounts of silt, partially decomposed tree litter, and carbonized wood. Layers in the lower half of the unit contain more silt and humus. Some of these lower layers are similar in texture and color to those found in the present topsoils of dryer meadow habitats sparsely vegetated by xeric sedges and grass. The upper materials of the middle unit, however, are of forest origin and contain abundant charcoal. These layers, up to 2 feet thick, are well-mixed and unstratified suggesting turbation by tree-root growth or rodents.

The most striking feature of the middle unit are well-preserved stumps and logs, particularly abundant near its top. The distribution and the speciation of these coniferous stumps is shown in figure 3-6 and tabulated in table 3-1. The uppermost stump line is associated with a two-foot forest layer that can be traced the full 2500-foot length of the main gully. The valley bottom was entirely forested at the end of middle-unit deposition. Figure 3-7 shows a 5-foot diameter stump of a Jeffery Pine that thrived during the latter part of middle-unit time. Several other pines reached 4-foot diameters. Jeffery Pine requires well-drained sites (U.S. Department of Agriculture, 1965). Wilde (1958, p. 207) indicates that water table depths of 5 to 7 feet are optimal for coniferous forest trees. It is reasonable to assume that the valley-bottom water table was deeper than 5 feet for the growth of these large pines in East Meadow. Water table depth is an important parameter of the valley bottom environment that will be used later in a discussion of hydrologic variations during the Holocene.

Chronology of the middle unit: A radiocarbon age of 2830 years B.P. (UGa-448) was obtained from the stump, numbered "A5", (figure 3-6), in the center of the middle unit. This is the root mass of a fir tree that grew in a soil layer just beneath, and truncated by, the uppermost soil layer that contains the prominent stump line. In a discussion of the chronology of the upper unit it will be shown that the death age of the uppermost stumps is about 2500 years B.P.

The length of time represented by the middle unit is not closely bracketed. At East Meadow, the interval must lie between 9020 years B.P. and c.a. 2500 years B.P. It is reasonable to ask whether the stratigraphic boundaries of this middle unit are really time

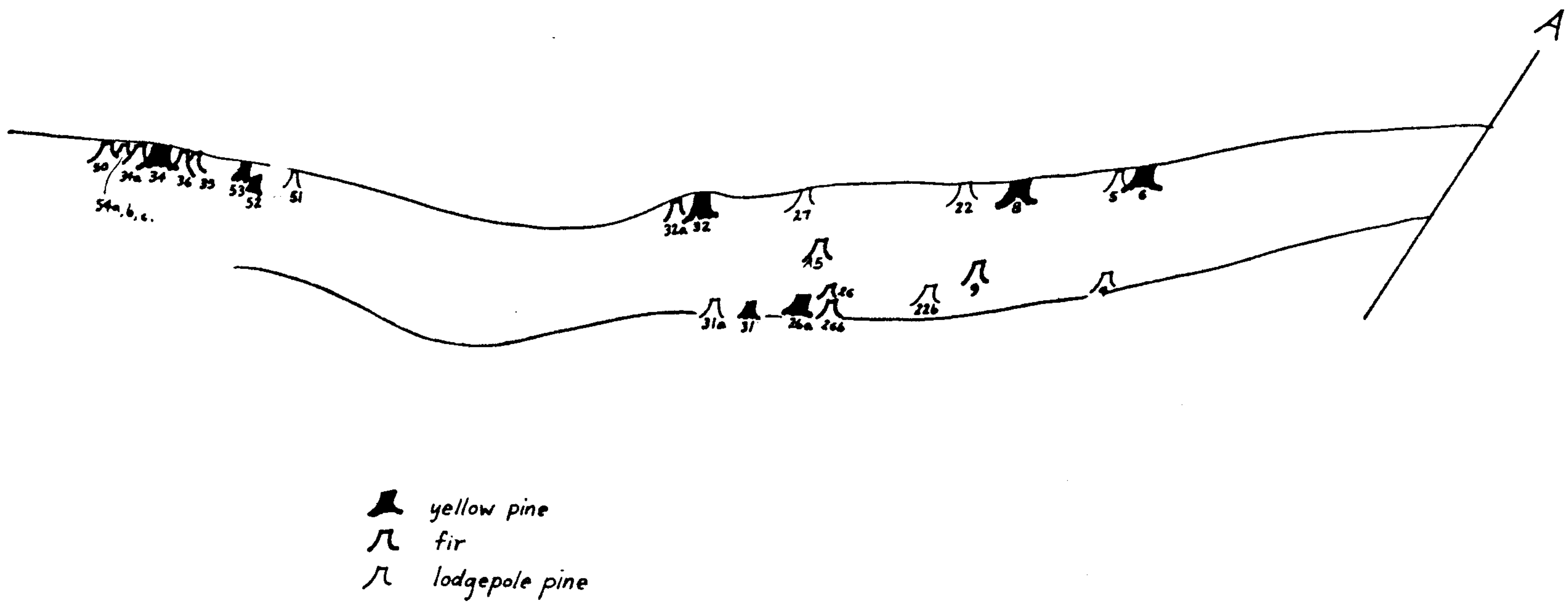


Fig. 3-6. Distribution and species of buried tree stumps (in situ) in middle Unit East Meadow, Yosemite National Park, Calif.

Figure 3-7. Stump no. 34 (table 3-1), a five-foot diameter yellow pine rooted in the top of the middle depositional unit. The surrounding soil and overlying meadow deposits have been entirely removed by erosion. A remnant of the meadow surface is behind the stump. East Meadow of Aspen Valley, Yosemite National Park.

Table 3-1. in situ stumps of conifers in the middle depositional unit, East Meadow, Yosemite National Park.

No. on Map	Stump diameter (inches)	Largest resin ducts* (microns)	Genus	Probable species	Remarks
5	36	100	Pinus	contorta	
6		200	"	jeffreyi	charred top
8	48	200	"	"	charred top
9	21	none	Abies	?	
22	24	100	Pinus	contorta	charred top
22b	20	130	"	"	
26	35	none	Abies	magnifica	red root bark
26a	14	200	Pinus	jeffreyi	
26b	20	none	Abies	?	
A5	24	none	Abies	?	C^{14} age = 2830 yrs.
31	20	none	Abies	?	
31a	20	100	Pinus	contorta	
32	42	150	Pinus	jeffreyi	
32a	10	none	Abies	?	charred top
34	63	200	Pinus	jeffreyi	charred top
34a	48	none	Abies	?	
35	36	none	"	?	
36	24	none	"	?	
50	40	none	"	?	
51	6	100	Pinus	contorta	
52	5	200	Pinus	jeffreyi	
53	6	200	Pinus	jeffreyi	charred top
54a	14	none	Abies		
54b	30	none	Abies		
54c	16	none	Abies		

* Comparative wood anatomy from Panshin, De Zeeuw, and Brown (1964, p. 443-482)

lines as has been tacitly assumed in the preceding discussion. The possibility exists that the major boundaries in the section represent encroaching environments as the valley aggraded through time. If a transverse section were available across the valley such relationships might be seen as the valley bottom environments onlapped the side slopes during accumulation of the fill. However, the conformable relation of the litho-stratigraphic boundaries of the major units with the internal bedding within the deposits strongly suggests that these are time lines within 50 to 100 years.

The valley bottom environment during middle-unit time was well drained. No deep channels are observed in the section that would suggest good drainage was provided by incision. Instead, it appears that aggradation prevailed during this time, the material of which being largely coarse pervious debris derived by slope wash. There may be depositional hiatuses in the 6500-year stratigraphic record of the middle unit, but detection of such hiatuses would require more detailed radiocarbon dating.

The dryer environment and deeper water table indicated by middle unit deposits can be explained by a much diminished supply of ground water caused by a change in the amount of precipitation or pattern of snowmelt.

The upper unit

The contrast between the stratified dark peats of the upper unit with the underlying lighter colored, crudely stratified forest materials of the middle unit produces everywhere an obvious contact which is also marked by the conspicuous line of stumps beneath. The upper unit thickens from 3 feet in the lower part of the meadow to 15 feet near the upper end as shown in figure 3-2. The character of the sediment layers also change up-meadow. Sediments in the lower reach are parallel-bedded, massive, sedge peat with intervening beds of well-sorted gravels, but headward the massive sedge-peat layers interfinger with, and grade into, sandy humic loam and sandy silt beds. The gravel layers are also siltier and less well-sorted up-meadow. Cut and fill cross-bedding relationships (amplitudes less than 1 foot) exist among the gravel and sand layers in the upper reaches in contrast to the horizontal bedding down-meadow. Individual gravel layers and the thinner peat layers are rarely continuous for more than a few hundred feet, except for the uppermost peat layers, which can be followed continuously for about 1500 feet along the trunk gully. It attains a thickness of 4 feet in the center of the meadow. Below this massive peat are usually two or three prominent gravel layers separating two thinner peat beds (figure 3-3).

The massive peat layers are composed of interwoven fibers and roots of sedge. Around the upper margin of the meadow they become increasingly silty, and the thinner peat beds are generally silt-rich throughout. Some contain considerable wood and bark fragments and conifer needles. All gradations from pure organic sedge peat through humic sandy loams (greater than 5 per cent organic by dry weight) occur in this upper unit of East Meadow. Both peat and the humic sandy loams appear dark brown in the section because of their organic content and their capacity to hold moisture. The organic material in these layers has not sufficiently decomposed to produce muck like that in the basal unit.

Volcanic Ash: Within the peat is one prominent and nearly continuous layer of volcanic ash varying in thickness from 0.4 to 1.5 inches. The layer is assigned a maximum radiocarbon age of 1545 years B.P. from analysis of a lump of charcoal found within it (Wood, 1972). It is identified as tephra 2 (Chapter 4) closely bracketed elsewhere by radiocarbon dates at 1200 years B.P. Thus the 1545 years B.P. charcoal had already aged as heartwood of a tree with considerable longevity or as reworked surficial material before being incorporated within the waterlain ash layer.

Rate of accumulation of the peat above tephra 2 does not exceed 2 feet over the last 1200 years or 2 inches per century. Peat between tephra 2 and the top of the middle unit is bracketed between radiocarbon based ages of 1200 B.P. and 2830 B.P., thus its accumulation rate cannot exceed 5 inches per century.

At one section in the center of the meadow, a second, very thin and whispy layer of fine ash lies 0.6 feet beneath tephra 2. The two peat accumulation rates provide an age estimate for this whisp of ash of 1400 to 1670 years B.P.

Diatom layers: Easily mistaken for volcanic ash are several discontinuous whitish, diatom-rich layers which contrast sharply with the dark peats. A 15-cm thick layer of nearly pure diatom tests lies at the base of the upper unit. It apparently accumulated in a ponded area as shown in the section B - E, figure 3-2. Elsewhere in the upper peat and humic silt loam deposits, diatom layers occur as thin, 1 to 2 cm thick, layers composed of mixed finely divided organic material and whole diatom tests. Several taxa of diatoms are relatively plentiful, but they have not been specifically identified. It is thought that concentrations of tests into discrete layers may represent intervals during which parts of the meadow became water-logged or actually ponded, or perhaps forest-fire-induced changes in nutrient supply and water chemistry caused a flowering of diatoms. Detailed study

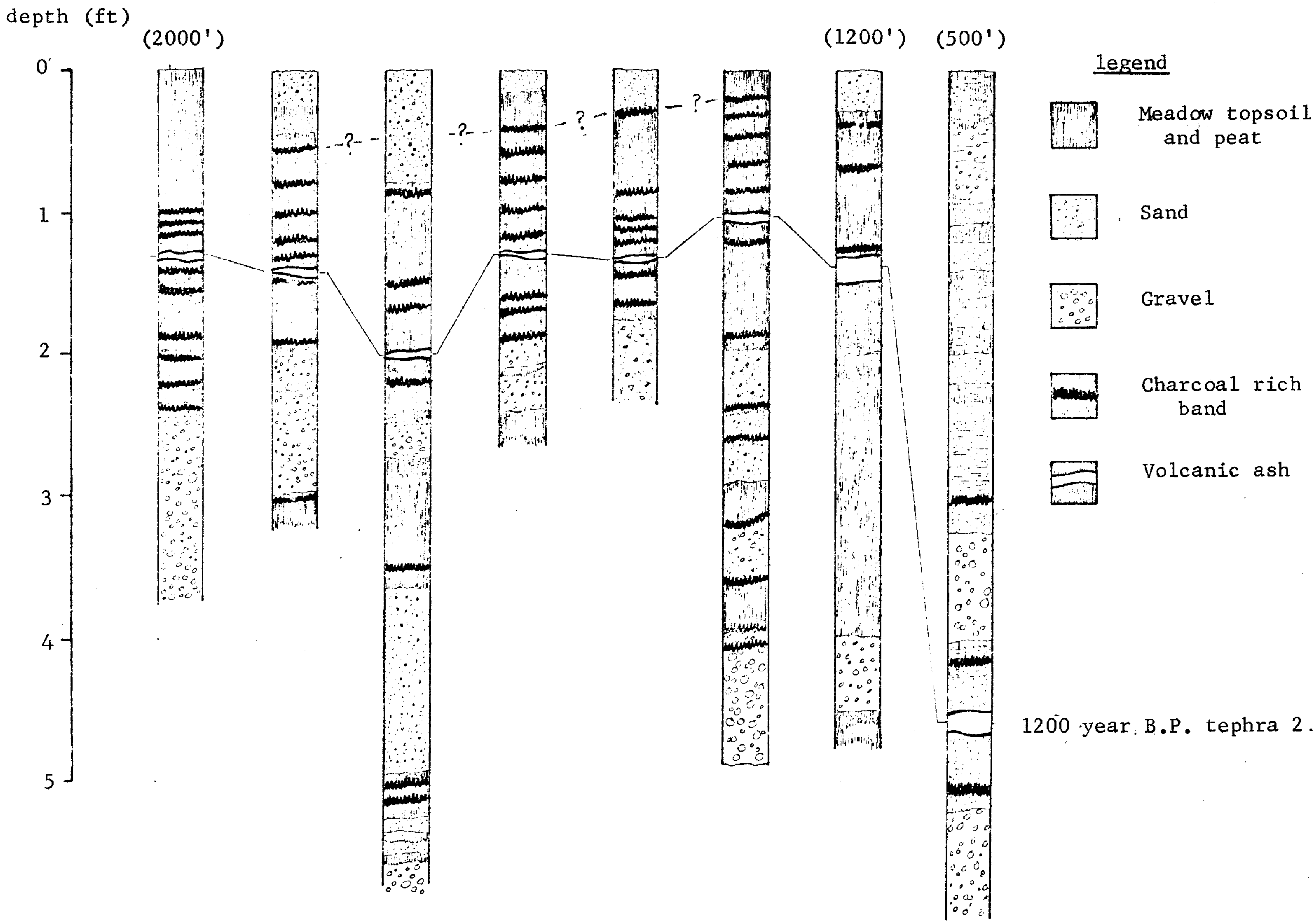


Figure 3-8 Detailed stratigraphy in the upper unit of East Meadow illustrating charcoal-rich bands above the 1200 year B.P. marker horizon. Numbers above columnar sections refer to horizontal distance west of point "A" on index map of figure 3-2.

of diatom stratigraphy might be useful in revealing subtle environmental variations during the accumulation of meadow deposits.

Charcoal bands: Thirteen separate charcoal-rich bands are found within peat layers of the upper unit. They are not continuous, but, assuming that these layers record major forest fires, it was hoped that each would serve as a time-correlative marker horizon. A 380-year fire history extending back from the present was determined from study of annual growth rings on recently logged, fire-scarred, sugar pine stumps that surround East Meadow. Charcoal bands within sections of the upper unit exposed along the main trunk gully are shown in figure 3-8. In four of the eight sections examined, exactly 5 bands are found above the 1200 year volcanic ash layer. In the remaining sections, 2 to 3 bands are found. This consistency is probably significant, but matching these layers to the known fire history is difficult because East Meadow fire history is imperfectly recorded prior to the late 16th Century. As shown in figure 3-9, fires were recorded on 3 or more stumps for the years 1779, 1803, 1842 and 1901 A.D. The latter 2 fires were also noted by Wagener (1961) in a similar study at Crocker's Station, 8 miles to the southwest. He considered only the 1842 and 1901 events to be major fires. Wagener found a prevailing 8-year natural fire frequency in the west slope Sierra timber zone which compares well with a 7 to 9 year frequency reported for the Sierra mixed Sequoia-conifer forest by Kilgore (1973).

Forest fires of this frequency are not recorded by charcoal layers. It is likely that only major fires, and perhaps, only those followed by a heavy slope-washing runoff, produce significant charcoal layers. If these layers represent major fires, then there have been five such events in the East Meadow watershed in the past 1200 years. It cannot be said with certainty which, if any, of the dendrochronologically recorded fires matches the uppermost charcoal layer in the section, but the thickness of peat capping that band would be consistent with the 1842 A.D. fire or one still older.

Gravel beds: Well sorted, sandy gravels derived from grus are interstratified with the peat. Three prominent gravels shown in figure 3-2, and figure 3-4, occur discontinuously as lenses up to one foot thick and several hundred feet long beneath the 1200-year marker ash. Two more gravel layers occur above the 1200-year ash in the upper reaches of the meadow producing thereby a thickening of the upper unit with the aid of intercalated sand lenses. These sand and gravel layers do not extend to the center of the meadow.

Clasts in the gravel lenses rarely exceed 1 cm diameter. The size analysis given in

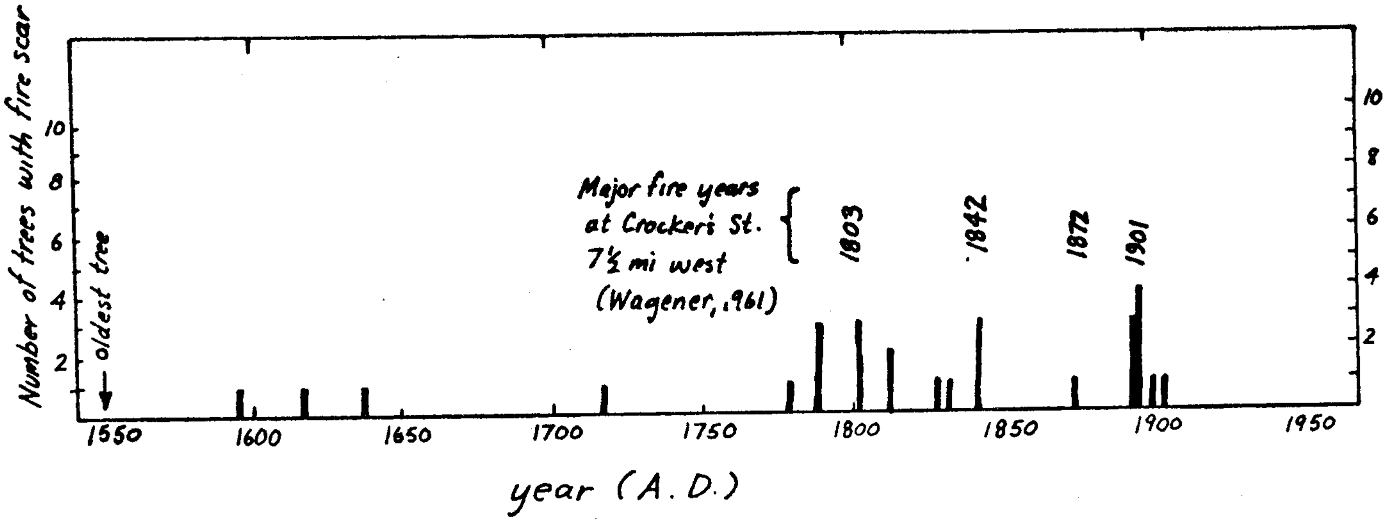


Figure 3-9. Forest-fire history at East Meadow of Aspen Valley, Yosemite National Park as read from fire scars on Sugar-Pine stumps. Uncertainty in dendrochronologic age of fire is probably ± 5 years for there is uncertainty of this order in the field tree-ring counts, and the logging operations spanned 1950 - 52 .

figure 3-10 is typical. Median size of the granules is 1 to 2 mm reflecting the texture of the granodiorite from which they are derived. The gravels are well-sorted with a Trask sorting coefficient Q_3/Q_1 between 2 and 2.2.

The pure gravels show little stratification, but sandier beds are crudely layered. In the upper reaches of the meadow, some of the sands display small scale cross-bedding less than 1.0 foot in amplitude associated with correspondingly small scale scour and fill.

Material comprising relatively large gravel bars along the smaller Sierra streams (figure 2-13) is similar in texture and stratification to the gravels in the East Meadow section. Jahnda (1967) reports that relatively large flat-topped gravel bars and berms of present High Sierra mountain streams are constructed during floods with recurrence intervals of more than 10 years. The gravel beds in the East Meadow section probably represent flat-topped gravel bars of this origin.

It is surprising that only 5 significant gravel beds were spread over the meadow surface during the past 2500 years. This indicates that gravel transporting flows have been relatively infrequent during the deposition of the upper unit within the meadow deposits. The recurrence interval for bar-building floods suggested by these five layers is about 500 years.

A 124-year historical record of Tuolumne drainage flood stages comes from La Grange (elev. 330ft) where the Tuolumne discharges into the San Joaquin Valley. Major floods occurred in January, 1862, January 1911, and December, 1955. The 1862 flood is considered the largest historical flood (Young and Cruff, 1967) in the Tuolumne drainage. Peak flow at La Grange is estimated by Wagoner (1908) to be 130,000 cfs or 85 cfs/sq. mi., but there are no data for higher elevation stations.

Since no gravel beds are found in a position in the East Meadow section that can be attributed to the 1862 flood, the five gravel beds in the section may be from flood flows greater than those of 1862. Were a flood of the intensity of the 1966 southern Sierra flood to occur at East Meadow a discharge of (3.5 mi^2 x 400 cfs/mi^2) 1400 cfs might flow across the meadow. It is suggested that the gravel beds in the East Meadow section were deposited by flows of this magnitude. According to Dean and Scott (1971) the 1966 southern Sierra flood produced peak discharges three times greater than the computed 50-year recurrence flood discharge. Hence it is not unreasonable to suspect that each significant gravel layer in the meadow sections represents a flood event of several-hundred-year-recurrence interval.

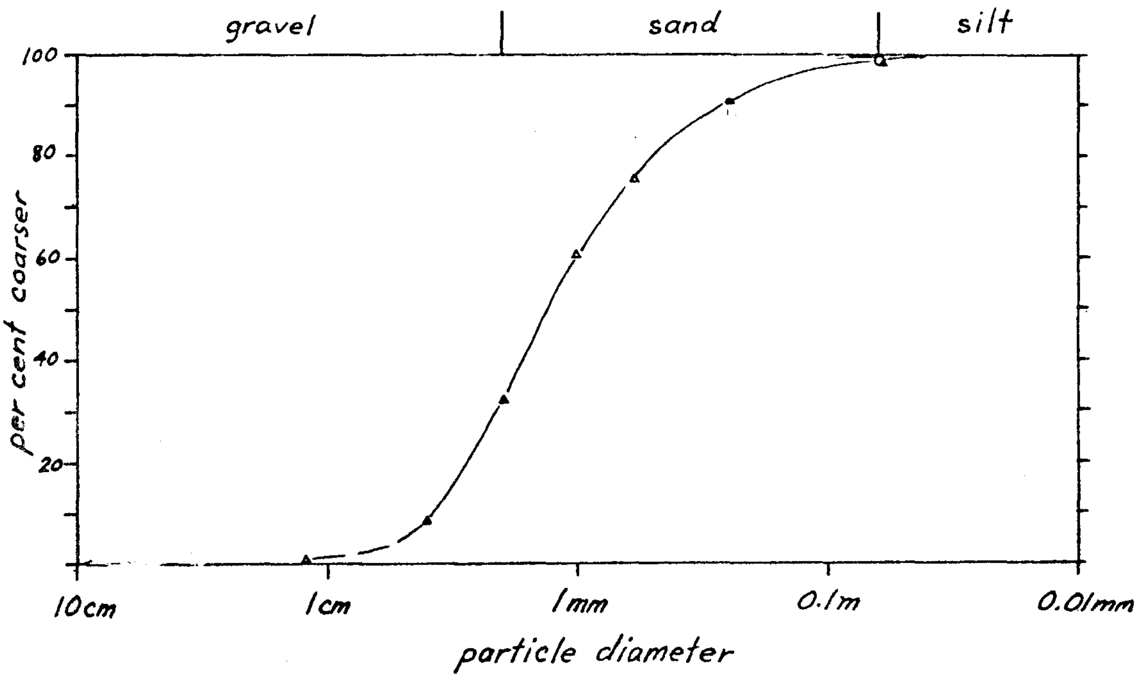


Figure 3-10. Size analysis of material in gravel bed immediately beneath the massive peat bed of the upper depositional unit, East Meadow.

Chronology of the upper unit: Ages within these deposits are established by; the 1200-year B.P. tephra 2, the surface of an historically-known meadow that began eroding in 1911 A.D., and the age of underlying deposits. The date of the contact representing the change from forest to meadow conditions lies between the 2830 year B.P. age on the st mp in the middle unit and the 1200 year B.P. tephra 2 layer. Radiocarbon dating the outer wood of any one of the many stumps would give a most accurate age, but it can as well be estimated without the additional cost. Since the 1200 year marker horizon is in the middle of the upper unit with perhaps slightly more peat beneath than above, the lower half represents an estimated 1300 years, giving an age of 2500 years B.P. for the total unit. Since the 2830 year B.P. stump is overlain by a layer of forest soil that contains an *in situ* 3-foot diameter pine stump, an interval of some 300 years seems reasonable. This date of 2500 years B.P. marks the death of the forest and the initiation of meadow conditions.

The contact between forest soils and the meadow topsoils indicates a rise of at least 5 feet in the shallow ground water table of the valley bottom about 2500 years ago. The yellow pine forest prior to its death required that the water table be at least 5 feet below the surface, but meadow conditions indicate a water table less than two feet.

Forest fire is implicated by the charred tops of 5 of the uppermost buried stumps of unit 2, table 3-1. But it will be shown later that forest fire alone cannot account for a 5-foot water table rise. The possibility that the water table rise was caused by damming or impeded soil drainage was considered. However, no vestige of any slides, log jams, or beaver workings are found below the meadow, and examination of the longitudinal profile of the valley, figure 3-2, shows that damming alone would not explain the poorly drained sloping surface. The water-table rise that initiated meadow-conditions in the valley bottom was caused basically by climatic variation and was possibly triggered by forest fire. This mechanism is discussed after consideration of the other sites.

Exchequer and Cabin Meadows, Sierra National Forest

Setting

Exchequer and Cabin Meadows lie in adjacent valleys separated by a 300 foot ridge

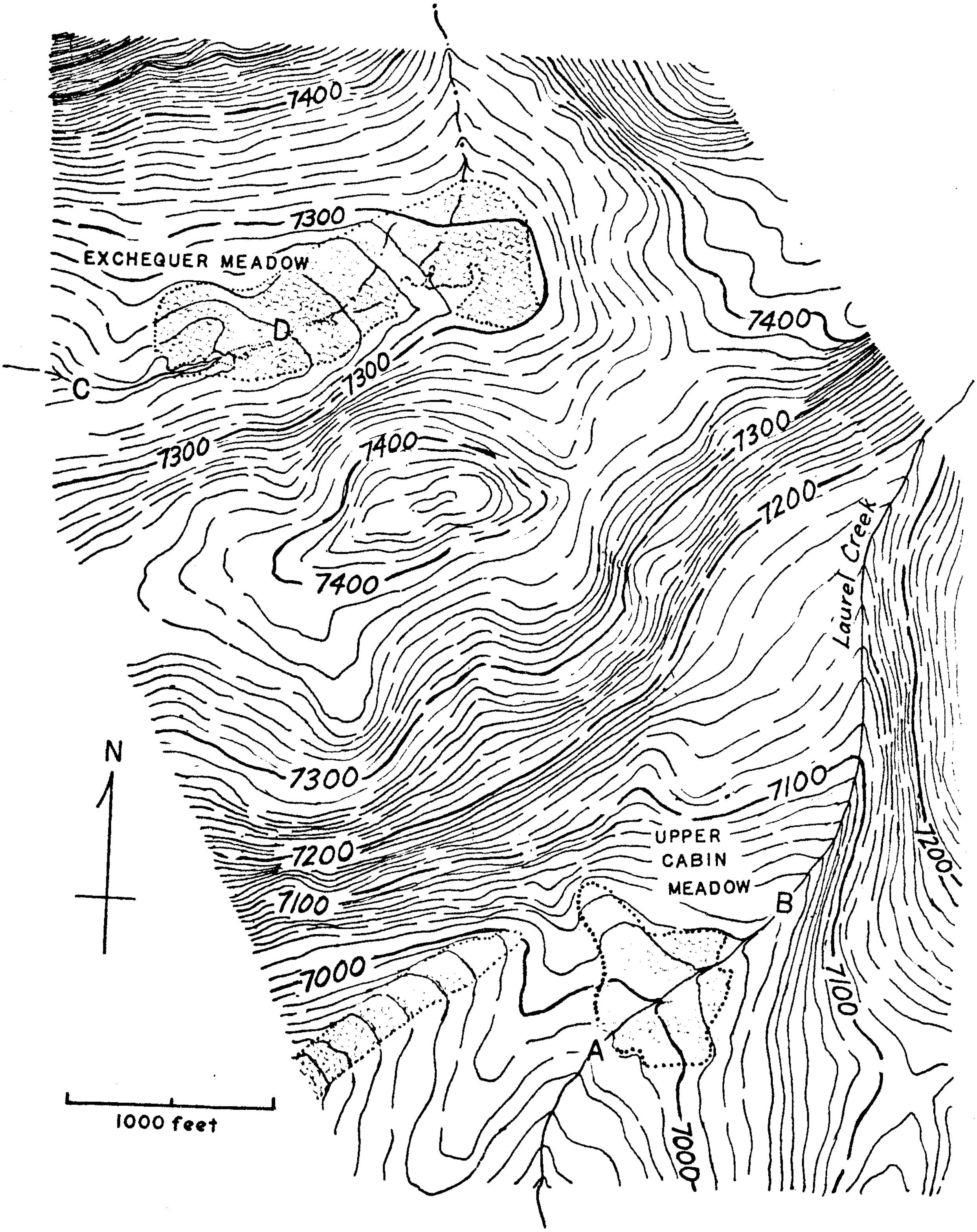


Figure 3-11. Location map of Cabin and Exchequer Meadows, Dinkey Creek area of Sierra National Forest showing sections A-B and C-D of figures 3-12 and 3-14 . 10-foot contours are from plane-table survey of meadows and interpolated between 80-ft contours on the Huntington Lake 15' quadrangle map elsewhere.

(figure 3-11). The valleys cross an upland tread at 7000 foot elevation that rises 1500 feet above the canyon of Dinkey Creek. Bedrock is hornblende-biotite granodiorite of the Dinkey Creek pluton (Bateman and Wones, 1972). Valley sides are covered with mixed-conifer forest of the main timber zone with a few black oak on south facing rocky slopes. These valleys were probably never glaciated, and certainly not during the Wisconsin, although several small Wisconsin cirque glaciers terminated on Bear Mountain, a few miles away, at elevations near 7300 feet as indicated by moraines mapped by Matthes (1960), and Bateman and Wones (1972).

Eighteen-acre Exchequer Meadow is roughly an 1800 by 400 foot rectangle. It is watered by seepage from a 0.7-square-mile watershed above the meadow including surrounding slopes and Exchequer Creek which is ephemeral above the meadow. Meadow vegetation is a sedge-rush-herb community described in Chapter II. Lodgepole pines 40 or more years old occur within the meadow as scattered clumps of three or four trees and as a 1-acre grove. Willow thickets locally line the stream courses. Exposures at Exchequer Meadow are presently limited to a deep discontinuous gully at the lower end, and a few shallow gullies in the middle.

Upper Cabin Meadow has been destroyed by very recent gullying causing a 10-foot drop in the water table. Young lodgepole pines now cover much of the dessicated, dusty flat that was once a 9-acre, roughly circular meadow. The former meadow was watered by perennial Laurel Creek and by drainage from surrounding slopes comprising a 1.1 square-mile watershed. A continuous gully exposes a complete longitudinal section through the meadow (figure 3-12).

Stratigraphic section, Upper Cabin Meadow

Laurel Creek has cut 10 to 15 feet downward through the fill beneath Cabin Meadow to a buried soil profile developed upon coarse alluvium (figure 3-12). The profile consisting of an O_2 horizon, A horizon, AC_g horizon, a C_g, and C horizon is well-preserved and illustrated in figure 3-13. Cobbles and boulders rich in mafic minerals in these horizon are disintegrated and can be cut with a shovel. Felsic cobbles are sound. Roots, charcoal and wood are well-preserved in the 1.5 foot thick O and A horizons. Contact with the underlying AC_g horizon is transitional over 2 inches. The greenish-grey (5GY 5/2) color of the AC_g horizon developed into the cobbly alluvium indicates that it is gleyed.[1]

footnote for page 66

[1]Gleyed horizon: Papadakis (1969, p. 123), in defining a horizon on the basis of color or iron state, requires that the gley horizon have a moist chroma of 4 or more, 1 or less if hue is 2.5Y redder, 2 or less if hue is 5Y or 7.5Y. Hues 10Y and bluer are entirely included in gley. The hues of the gley horizons found in the paleosol meet the latter criterion for gley.

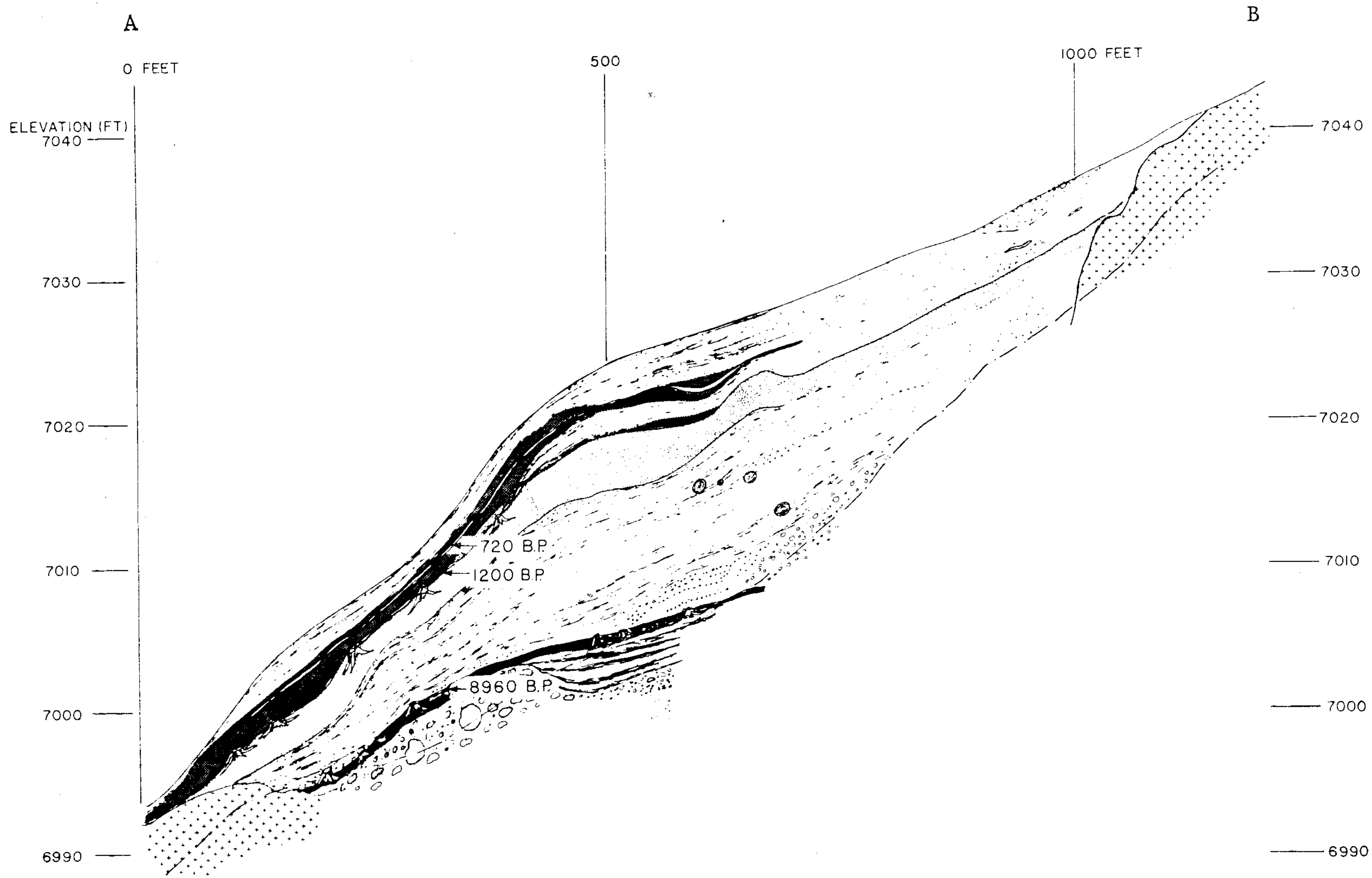


Figure 3-12. Longitudinal section through Upper Cabin Meadow on Laurel Creek, Dinkey Creek Basin, Sierra National Forest, California. Refer to legend on figure 3-2, page 49. Vertical exaggeration = 15 x.

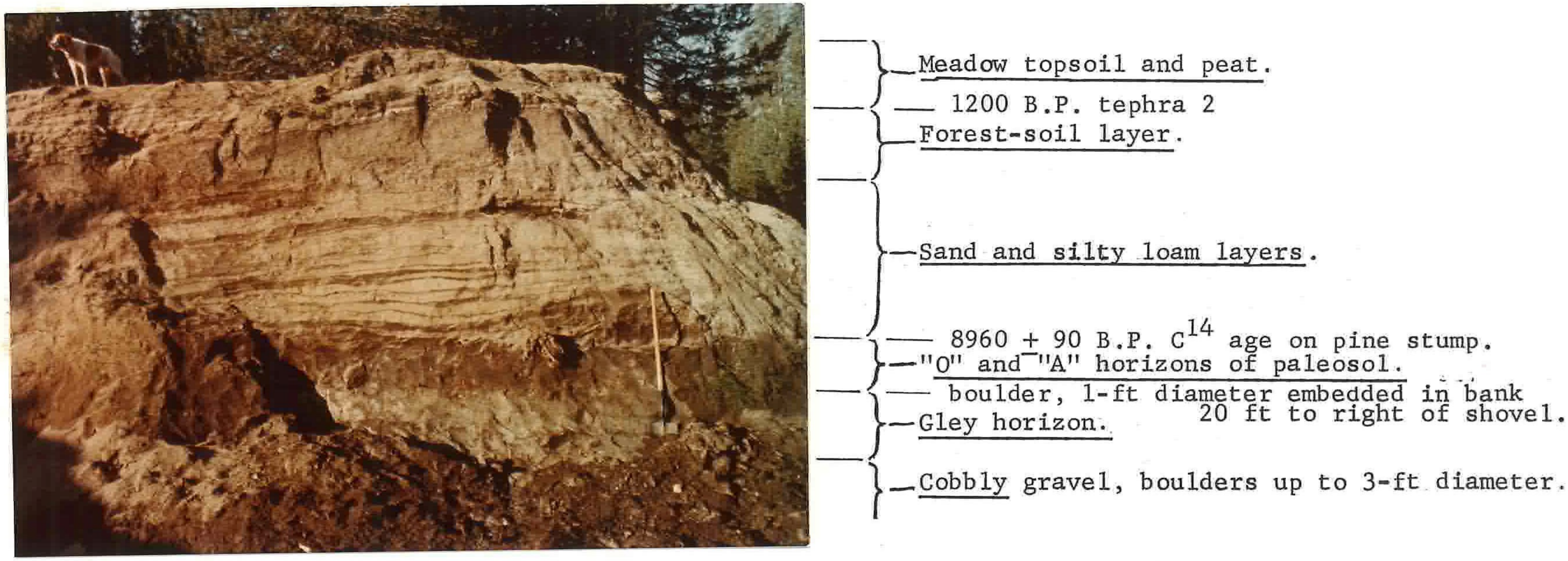


Figure 3-13. Stratigraphic section at Upper Cabin Meadow on Laurel Creek, Dinkey Creek Basin, Sierra National Forest. Photo illustrates the gleyed nature of the basal paleosol and location of radiocarbon date UGa-621. Shovel and handle are 6 ft long. Digging spoils in foreground are not part of the section.

There is considerable clay and silt in the gley horizon, either from illuviation or *in situ* decomposition of parent alluvium, making it tough and erosion resistant. The paleosol line can be traced for 600 feet along the creek bed and is marked by well-preserved stumps of lodgepole and willow. The radiocarbon date of 8960 years B.P. is from wood of a small willow stump lying on its side partly embedded in the "O" horizon at the very top of the paleosol (figure 3-13). An auger hole, started in the stump line, 250 feet upstream from the last exposure of the gley horizon, penetrated 4.5 feet of silty muck and fine sand layers, and 3.5 feet of inorganic sand and gravel. This suggests that while soil was forming upon an elevated bar of alluvium, fines were accumulating in a willow swamp or wet meadow in a swale a few feet lower in elevation. The stream gradient at the paleosol stratigraphic level was 1.7 per cent for the upper 400 feet of exposure steepening to 3.3 per cent for the lowermost 120-foot reach which merges with the bedrock channel below the meadow.

Deposited upon the paleosol is a 4 to 10 foot thick sequence of alternating layers of poorly sorted sands, fine gravels, and silt-loam soil. This sequence is overlain by a 1 to 2.5 foot thick unstratified, coarse, sandy loam with abundant charcoal, regarded as a forest soil layer. At the top of this layer are several truncated stumps in growth position. Some are charred but otherwise preserved and still retain a pine oder. All are identified from resin-duct size (Panshin and DeZeeuw, 1964) as Lodgepole Pine. Thickness of the forest soil layer indicates that a forest covered much of the valley bottom for a considerable length of time, but this interval is not radiocarbon dated. Resting immediately upon the stumps, but really constituting part of the overlying peat layer are thin lenses of pumiceous volcanic ash identified as the 1200 year B.P. tephra 2 (figure 3-13). This marks the apparent abrupt transition to open meadow conditions. The overlying silty sedge peat is a 1 to 2 foot thick layer that grades upward into several interstratified layers of humus-rich sand loam and well-sorted fine gravel. The 720 year B.P. tephra layer occurs in the peat layer, and provides a chronological line for the entire length of the section.

Stratigraphic section, Exchequer Meadow

The basal material at Exchequer meadow is also a coarse alluvium (figure 3-14). The alluvium is overlain by a black organic silt layer 0.5 to 1 foot thick containing many sticks

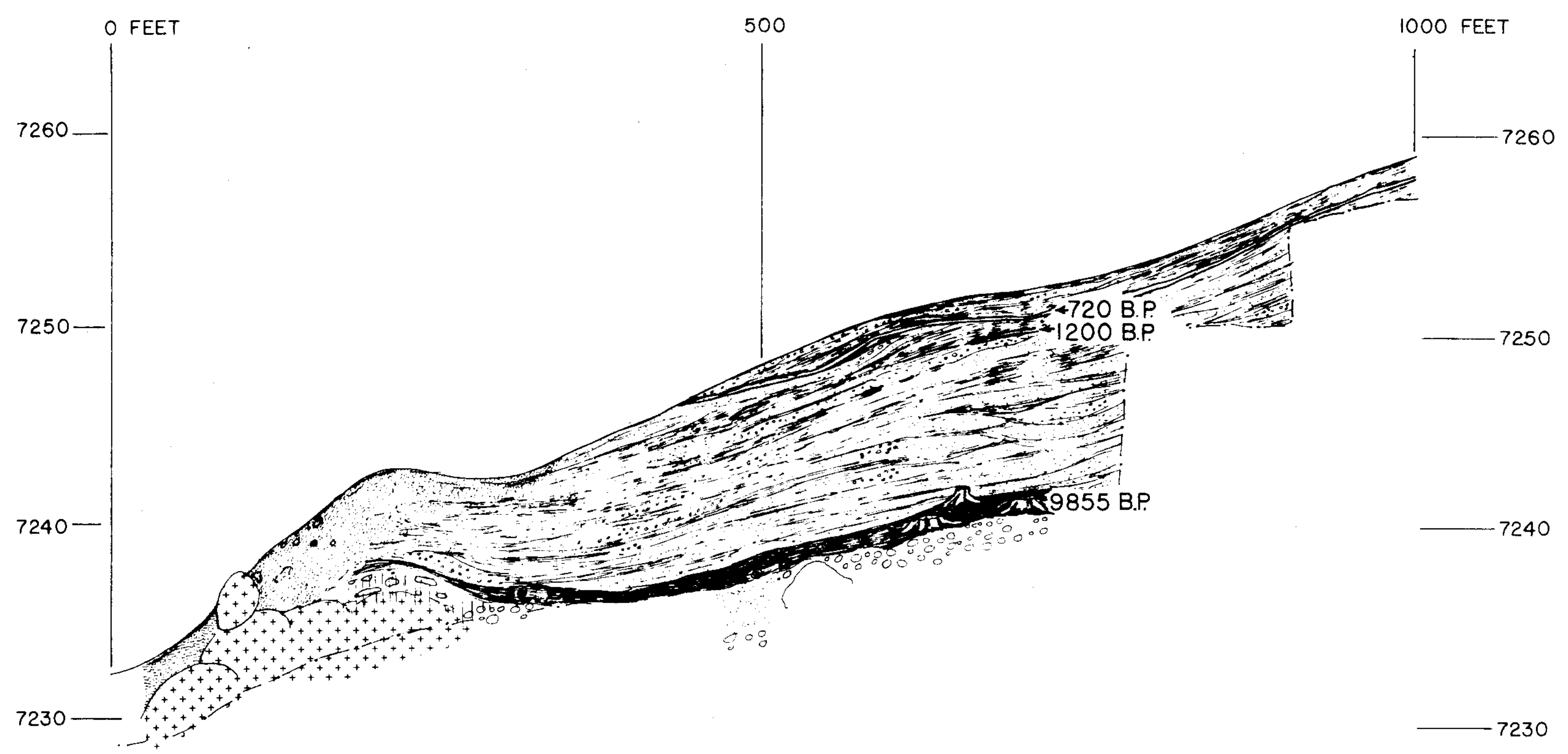


Figure 3-14. Longitudinal section through lower part of Exchequer Meadow, Dinkey Creek Basin, Sierra National Forest. Refer to legend on page 49. 15 x vertical exaggeration.

and twigs, Stumps are rooted on the top of this alluvial layer and within the black silt layer. They are truncated where the roots branch indicating that the above ground part of the stump rotted away before burial. Five of the stumps are identified as Lodgepole and Yellow Pine, the largest of which is radiocarbon dated at 9855 years B.P. The alluvium just beneath the organic silt is slightly cohesive and has a faint bluish hue suggesting incipient gleying. The deeper alluvium is unaltered.

Overlying the stump line are layers of sand and fine gravel alternating with humus-rich silt and sand-loam. This sequence continues to the surface indicating that this part of the valley has been meadowed for most of the Holocene.

The 1200 year B.P. tephra 2 layer appears on the walls of the lower gully, 1.5 to 3 feet beneath the meadow surface, as discontinuous lenses of white ash a few millimeters thick. The 720 B.P. tephra 1 layer occurs throughout the meadow as a 1 to 3 cm layer of ash and fine pumiceous lapilli, about 1 foot below the current meadow surface.

Chronology and interpretation

Basal sections are similar in both meadows. The coarse alluvium is presumed from the radiocarbon ages in the overlying soil to be pre-Holocene in age. Differing ages on wood within the similar paleosols suggest that soil formation may have occurred for at least 900 years. Gleying and weathering of the parent alluvium is far more extensive at Upper Cabin Meadow. The thin topsoil layers alternating with alluvium overlying the paleosol are indicative of moist, but not wet, meadow conditions. These conditions have persisted at the exposed site at Exchequer Meadow from c.a. 9000 B.P. to the present, but in Cabin Meadow, there was an intervening interval of forest cover. In future years, continued erosion of exchequer Meadow may uncover similar forest layers, but dissection to date has revealed only moist meadow soils, suggesting that this part of Exchequer Meadow was not affected by presumed hydrologic changes that allowed a forest to cover Upper Cabin Meadow prior to 1200 years B.P. Charred stumps at the top of the forest soil at Cabin Meadow suggest fire as a triggering mechanism for the abrupt transition to a peat forming wet meadow, but it will be shown that fire alone cannot be responsible for the significant water-table rise required to maintain the meadow.

Beasore Meadow, Sierra National Forest

Setting

Eighty-acre Beasore Meadow lies at 6800 feet in an upper tributary valley of Chiquito Creek Basin that heads against the southeast trending Chiquito Ridge, a spur of the main Merced-San Joaquin river divide. Mixed-conifer forest and small meadows cover most of the 1.5 square mile watershed above the meadow. Bedrock in the drainage is quartz monzonite and granodiorite (Huber, 1968). No glacial deposits are found within the drainage area; although Matthes (1960, Plate I) shows a morainal crest of Wisconsin age crossing the middle of the meadow. Jahnda (1966, p. 135) reports that he was not able to recognize such a moraine and further states that the extent of bedrock weathering in the drainage precludes Tahoe-age glaciation. In an adjacent drainage, 1.5 miles southwest of the meadow beneath a higher part of Chiquito Ridge, morainal deposits at Chilkoot Lake indicate both Tioga and Tahoe cirque glaciation. These glaciers terminated below 7200-foot elevation. On this basis, Jahnda (1966) estimates a local Wisconsin firn limit at about 7600 feet, which suggests that the Beasore Meadow drainage, although above timberline in Wisconsin time, was not likely to accumulate perennial snow sufficient for a glacier.

Beasore Meadow has a gently sloping, smooth surface about 200 feet wide that extends for nearly a mile along the valley bottom. A perennial stream flows through the meadow gathering rivulets of surface seepage and snowmelt from the valley sides. The upper meadow reach is somewhat boggy throughout the year. Vegetation is dominated by several species of sedge, but thickets of willow grow along stream courses and in clumps in the boggy upper-reach.

The middle 1500-foot reach of the meadow is dissected to a depth of 18 feet by discontinuous gully that in the summer of 1974 was extending headward into the upper meadow at point 'A' on figure 3-15. This dissection, and also that of nearby 25-acre Muglar Meadow, began in the late 1940s according to recollections of local residents.

The Stratigraphic section

The 12 to 18 feet of fill beneath Beasore Meadow is comprised of three stratigraphic units; 1) a basal sandy cobble-rich alluvial layer, at least 3 feet thick, 2) a 2 to 4 foot

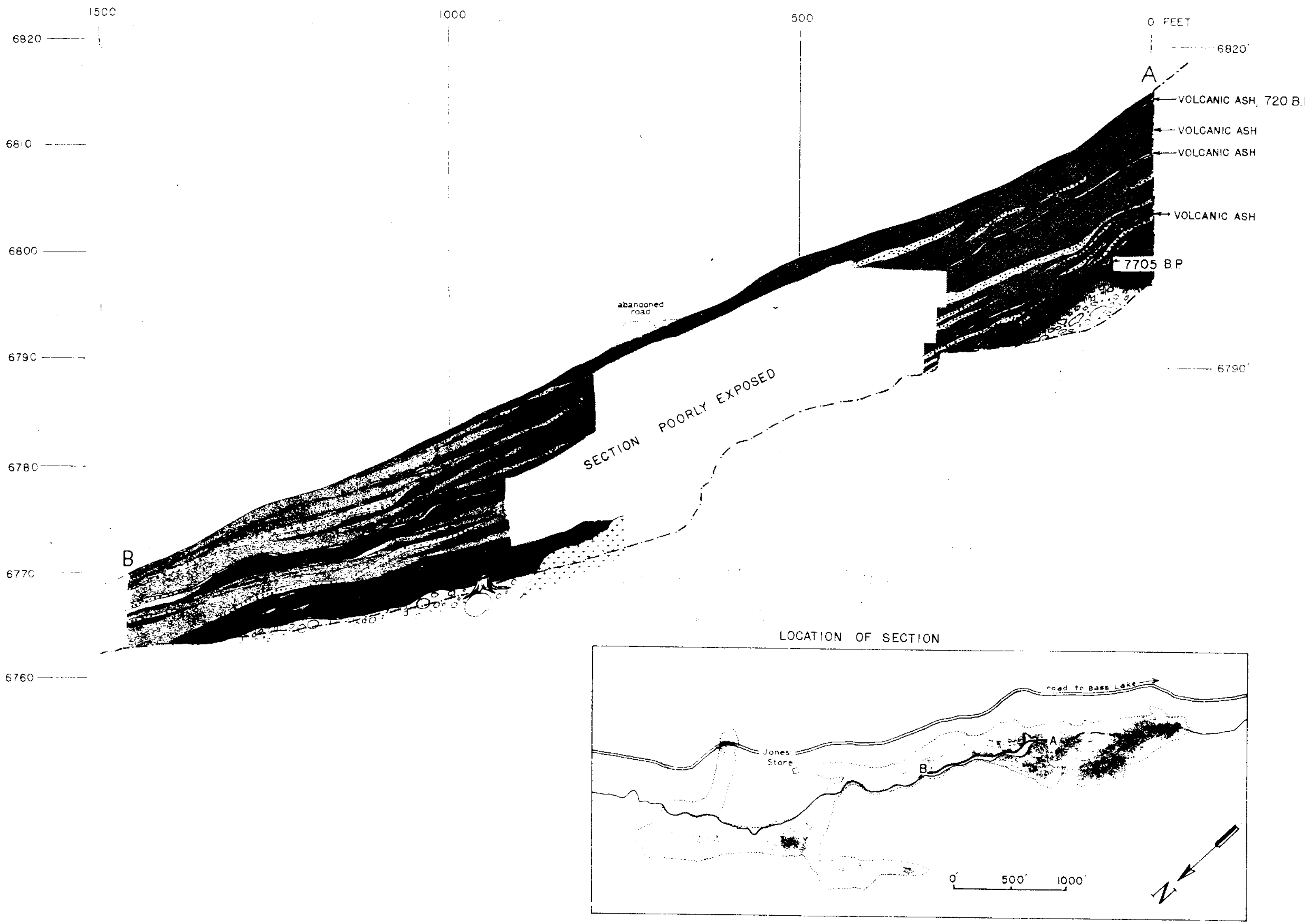


Figure 3-15. Longitudinal section through the central part of Beasore Meadow, Sierra National Forest. 15 x vertical exaggeration.

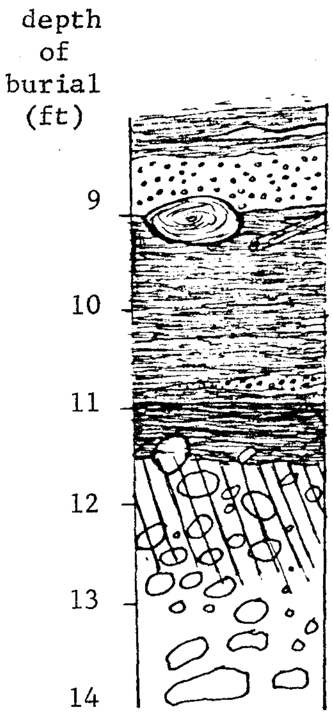
layer of decomposed, partly woody peat resting upon a gleyed paleosol profile, and 3) 8 feet of little decomposed sedge and willow-root peat which contains several fine gravel layers in its lower part. The 1500-foot longitudinal section charted from the walls of the gully (figure 3-15) illustrates the stratigraphic succession.

The basal sandy-cobble bed is exposed at both the headcut and in the lower reach of the gully where it forms the present stream bed. The cobble bed presumably lies upon bedrock. Larger clasts are sub-angular cobbles and broken exfoliation spalls of granitic rock. A few boulders up to 3 feet in diameter from this cobble layer lie in the stream bed. A grey-green gley horizon has developed about one foot into the cobble bed (figure 3-16). The 0.5-foot thick, black organic silty "0" horizon contains cobbles that extend into the underlying gley horizon. Two to four feet of dark brown, fine, decomposed, silty peat overly the "0" horizon. Abundant woody material occurs toward the top of this peat. A 1-foot diameter fir log of 56 annual rings lies embedded in the top of this peat. This log yields a radiocarbon age of 7705 ±90 years B.P.

Above the paleosol and peat are interstratified layers of sorted gravel and little decomposed sedge and willow-root peat. Fewer clastic layer occur in the uppermost 6 to 8 feet of mostly massive peat. This thick peat is probably representative of the section beneath the wettest part of the meadow. At the edges of the meadow, the peat grades laterally into humus-rich sandy loam and merges with coarse colluvial soils of the valley sides. In the massive peat at the headcut of the gully (point "A", figure 3-15), are four volcanic ash layers. The uppermost contains sandine microphenocrysts diagnostic of tephra 1, 720 radiocarbon years B.P., described in Chapter IV. The second layer is barely discernible being expressed by a few thin discontinuous lenses about 2 mm thick at most. The third and fourth layers can be traced along the walls of most of the gully, and are a few centimeters thick. Estimated ages are provided in the next discussion.

Chronology and interpretation of the section

The only firm dates on the Beasore section are 720 years B.P. on the uppermost ash layer and the radiocarbon age of 7705 years B.P. on the fir log buried 9 feet beneath the meadow surface. Discounting clastic layers, these indicate an accumulation rate for the massive peat between 3.6 and 4.2 cm per century. This rate divided into the range of burial depths for the other three volcanic ash layers yields peat-accumulation ages within



PROFILE DESCRIPTION

Fir log, C^{14} age = 7705 ± 90 B.P.
Dark brown (10YR 2/2) silty, decomposed peat containing abundant woody material at top.

"O" horizon. black (10YR 2/1) organic silt.

"G" horizon: grey-green (5GY 5/2) sandy silt matrix, slightly sticky, and granitic cobbles, some of which have disintegrated.

"C" horizon: granitic cobbles, gravel, and sand, sub-angular, unaltered, but partly iron stained.

Figure 3-17. Description of paleosol at Beasore Meadow, Sierra National Forest, California.

2230 to 2980, 3020 to 4910 years, and 6620 to 7705 years for the other three ash layers, respectively.

The basal paleosol appears similar to that found in the other meadows and its "0" horizon would presumably yield an age between 8700 and 10,185 B.P. The overlying peat and dated fir log may represent a somewhat different hydrologic environment than that which formed the paleosol.

The sedge and willow peat indicate a wet meadow established here shortly after 7705 B.P. The uniformity of the peat also suggests that this particular site in the meadow was not particularly responsive to variations in the hydrology which affected other meadows.

Wishon Reservoir Area, North Fork of the Kings River, Sierra National Forest

Several meadows lie in a 3-mile wide embayment bordering the gorge of the North Fork of the Kings River west of Wishon Dam. Bedrock mapped by Bateman and Wones (1972) is largely granodiorite. Glacial deposits from lobes of ice that extended west from the North Fork valley glacier modified drainage in the embayment, and most meadows lie in swales between lateral moraines (figure 3-17).

Reconnaissance glacial geology

Glacial deposits in the embayment were briefly examined in order to compare them with the better studied sequences in other west-slope drainages of the Sierra (Birman, 1963, Jahnda, 1967). Within the embayment are seven nested, morainal, arcs outside of which are thin till deposits mantling bedrock slopes to 150 - 300 feet higher than the outermost moraines (figure 3-17). Well-defined crests on moraines suggest a Wisconsin age. The inner 5 morainal arcs have the bouldery surfaces and sharp crests typical of west-slope moraines of the later Wisconsin (Tenaya and Tioga) glaciations. The outer two arcs have total relief of 5 to 20 feet and are not sharp crested. Boulders on their surfaces greater than one foot in diameter are spaced 30 or more feet apart. Between boulders the soil is composed largely of gruss. These outer two moraines, as well as the low moraines bordering Tule Meadow, the meadow on the west fork of Long Meadow Creek, and Sawmill Flat, may be as old as Tahoe. Till beyond identifiable moraines is probably pre-Wisconsin in age.

Stratigraphy beneath Hall and Tule Meadows

Since Tule and Hall Meadows lie in swales formed by the two outermost morainal arcs, it seemed worthwhile to auger them to determine the stratigraphic sequence. The stratigraphic relationships inferred from the augerhole in Hall Meadow are illustrated in figure 3-18. Three to four feet of stratified humus- rich sand and silt loam topsoil with thin sand lenses was found beneath the sod surface of both meadows. The 720 year B.P. tephra 1 was encountered 0.5 feet deep in both meadows. Beneath the 4-foot organic layer of both meadows, to a depth of six feet are layers of inorganic greenish-gray (5GY

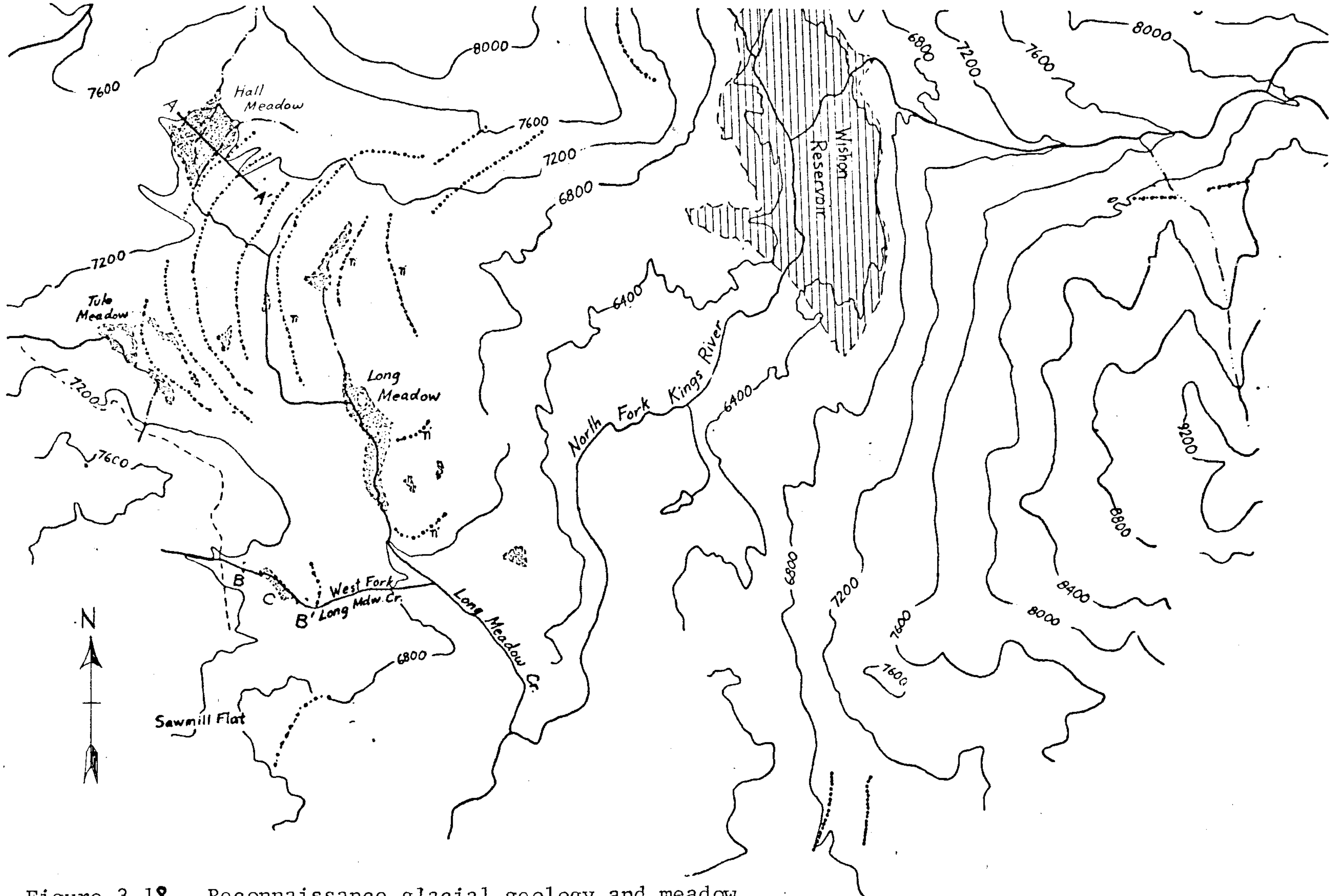


Figure 3-18. Reconnaissance glacial geology and meadow location map west of Wishon Reservoir, Sierra National Forest. Dotted lines are crests of moraines, Ti = Tioga or Tenaya age moraine. Dashed line is the upper limit of pre-Wisconsin ? till in the southwest corner of map.

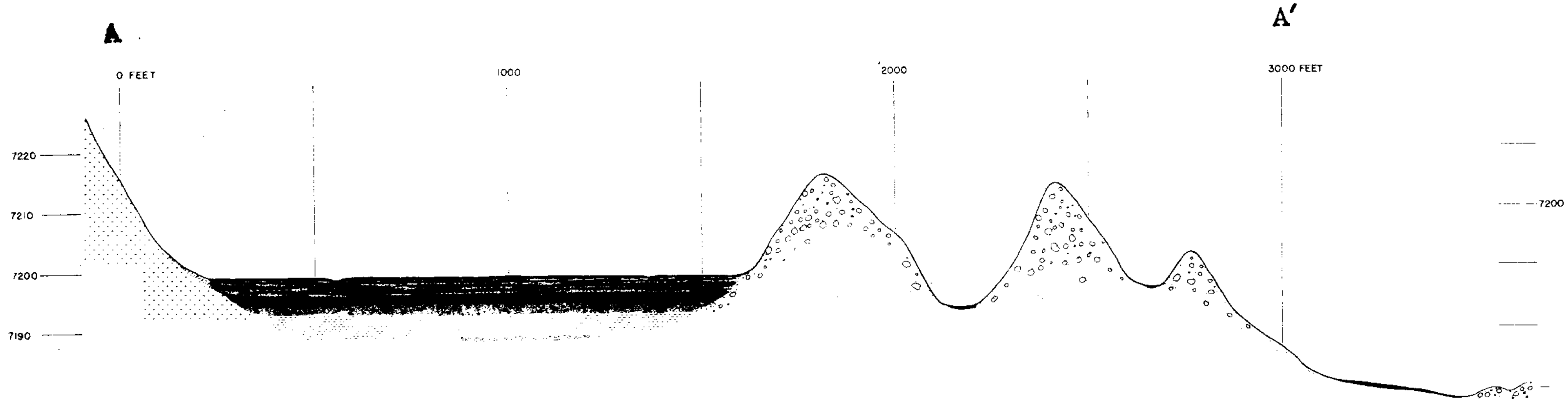


Figure 3-18 Transverse section through Hall Meadow, Sierra National Forest. Stratigraphy is based upon a 12-ft auger hole 1000 ft southeast of point A (figure 3-17). Glacial moraines are possibly of Tahoe age(discussed in text).

5/1) fine and coarse grained sands. In Tule Meadow, sand and gravel extend to 9 feet where pebbles over one inch prevented further penetration. Two thin peat bands 0.5 inches thick underlie Tule Meadow at depths of 4.5 and 7.5 feet. All other material below 4 feet in Tule Meadow was inorganic.

Beneath Hall Meadow, bluish-gray silt with scattered rootlets and possibly blades of sedge were encountered in the 6.4 to 7.4 foot interval. It is doubtful if sufficient sedge debris could be collected from this zone to yield a meaningful radiocarbon age for this deeper occurrence of organic material but the presence of organic debris is notable. From 7.4 to 9.7 feet is bluish-gray inorganic silt, composed of 0.5 to 1.5 mm laminae. This laminated silt is probably a rock flour from glacial meltwater deposited in quiet water. If the laminae should happen to be annual varves, then the 2 feet of 0.5 to 1.5 mm thick laminae may represent 400 to 1200 years of time. Ice standing behind any of the outer 4-morainal arcs could have caused ponding in Hall Meadow and diverted the drainage to the west. The laminated silt correlates with morainal deposits of a maximum advance probably of latest Wisconsin age (Tioga) and not with the more weathered outermost morainal arcs of probable Tahoe age. At a depth of 9.7 feet, artesian watererupted 1 ft. above the auger hole which was plugged and backfilled with considerable difficulty. The artesian water implies a permeable, and therefore sorted, clastic layer capped by the impermeable silt.

Thickness of the meadow deposits above the silts and sands is comparable with dated Holocene accumulations encountered elsewhere in the Sierra by Adam (1965) and Batchelder and Faverty (1972) in similarly formed depressions adjacent to receding glaciers.

Stratigraphic section, West Fork of Long Meadow Creek

A six acre meadow along this creek has been destroyed by a continuous gully that is in places 20 feet deep. The meadow was about 200 feet wide and 1500 feet long. The downstream reach of the gully has cut into underlying glacial deposits of unknown age. The upper reach of the gully bottomed on a resistant layer containing a jumble of large conifer logs embedded in layers of gravel and woody peat. The succession is charted largely from the south side of the gully (figure 3-19). It shows glacial till of probable early Wisconsin age overlain by 12 feet of woody peat and interstratified poorly sorted sands and fine gravels. Above the woody peat are several interstratified coarse sand and

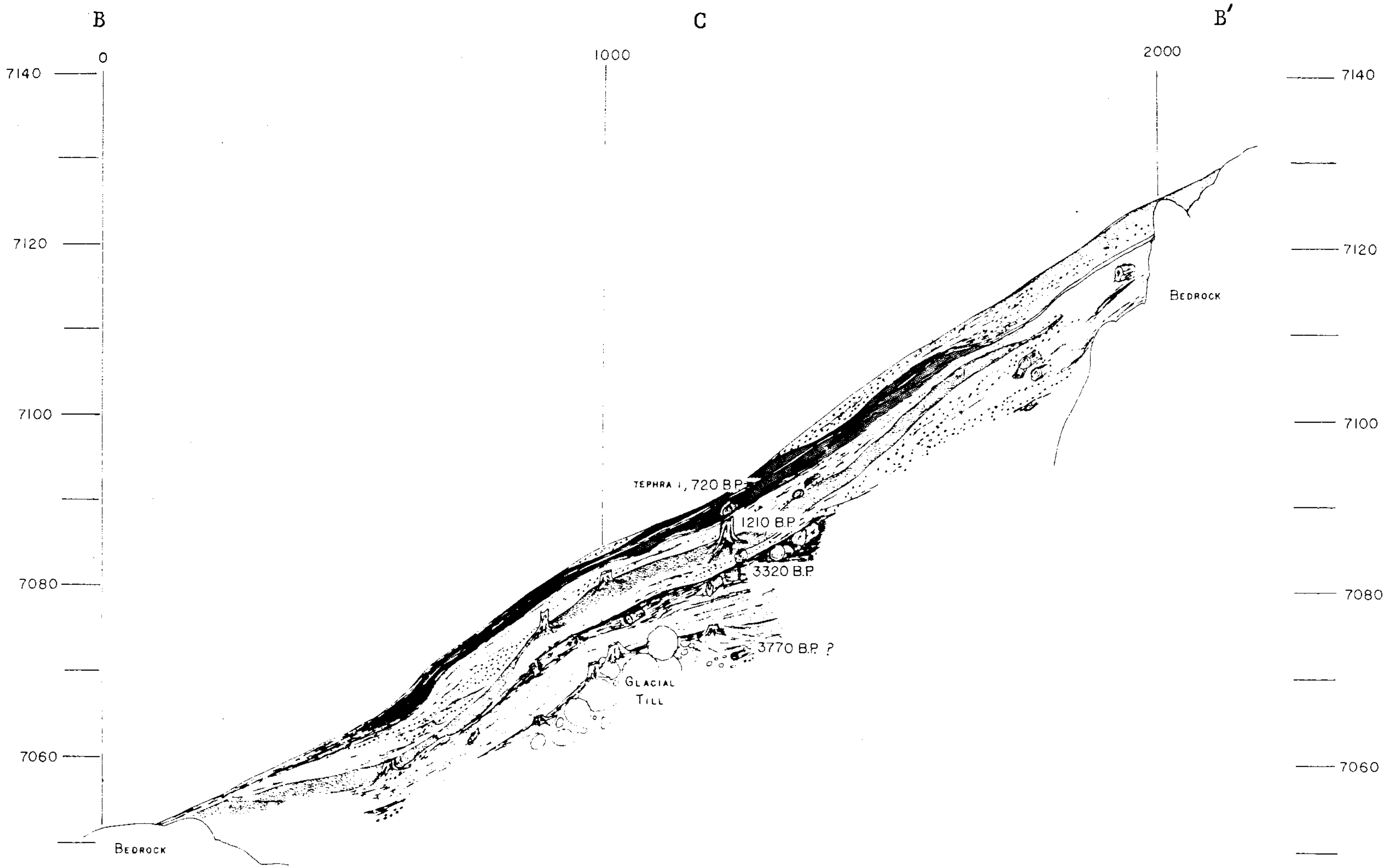


Figure 3- 19. Longitudinal section through meadow on tie west fork of Long Meadow Creek, Wishon Reservoir area of Sierra National Forest. Refer to legend on page 49. 15 x vertical exaggeration

silty peat layers which are overlain by a 2-foot thick forest soil layer. This soil layer contains several stumps of fir and lodgepole pine, some of which are rooted in the forest soil or the overlying coarse sand layers. The uppermost 2.5 feet of section contains mostly humus-rich sand-loam deposits and peat which are indicative of open meadow conditions. Figure 3-20 is a transverse section of the meadow deposits provided by a curve in the gully. It illustrates the difficulty in reconciling the stratigraphic levels in the lower part of the section on the north bank with the deepest radiocarbon age on the southbank. The north bank contains a basal paleosol with a thick gley horizon extending into the underlying till. At the top of the paleosol is a charcoal-rich organic-silt horizon similar in most respects but thinner than the "O" horizon of the basal paleosol in other meadows which has yielded radiocarbon ages between 10,185 and 8700 years B.P. Projection horizontally of the top of this paleosol intersects the south bank 5 feet above a radiocarbon dated log of 3770 years B.P. Radiocarbon ages in the south bank indicate a 10-foot accumulation of woody peat and poorly sorted sand within 430 years. The radiocarbon date is considered valid, and indicates that middle Holocene deposition may be complicated by episodes of scour and fill. No infilled gullies or obvious depositional hiatuses are seen in the sections. Identification of major scour and fill episodes in the depositional history of the middle Holocene unit would require more radiocarbon ages.

Above the 3330 year B.P. dated fir log, the section on both sides of the gully can be reasonably matched (figure 3-21). Two radiocarbon ages were determined on the outermost wood of a pine stump and a horizontal fir log in order to bracket the age of a lens of volcanic ash which was deposited upon the stump, but beneath the log. The age determinations agree within 35 years and are the basis for assigning an age of 1200 years B.P. to the tephra 2 ash layer in the southern Sierra. These ages also date the transition from a moderately well-drained site with scattered conifer trees to a wet meadow about 1200 year B.P. Wet meadow deposits are represented by humus-rich sand loams and silty sedge peat above the 1200 year ash layer. These deposits contain the 720 year B.P. tephra 1 layer which is here about 3 cm thick and can be traced the length of the gully.

The valley fill beneath the meadow is Holocene. The underlying till is probably early Wisconsin in that it is at the level of Tahoe-age moraines across the North Fork gorge (figure 3-17). In the Wisconsin, this till was dissected. In the Holocene, the valley began to aggrade. The age and stratigraphic relationships of the base of the valley fill

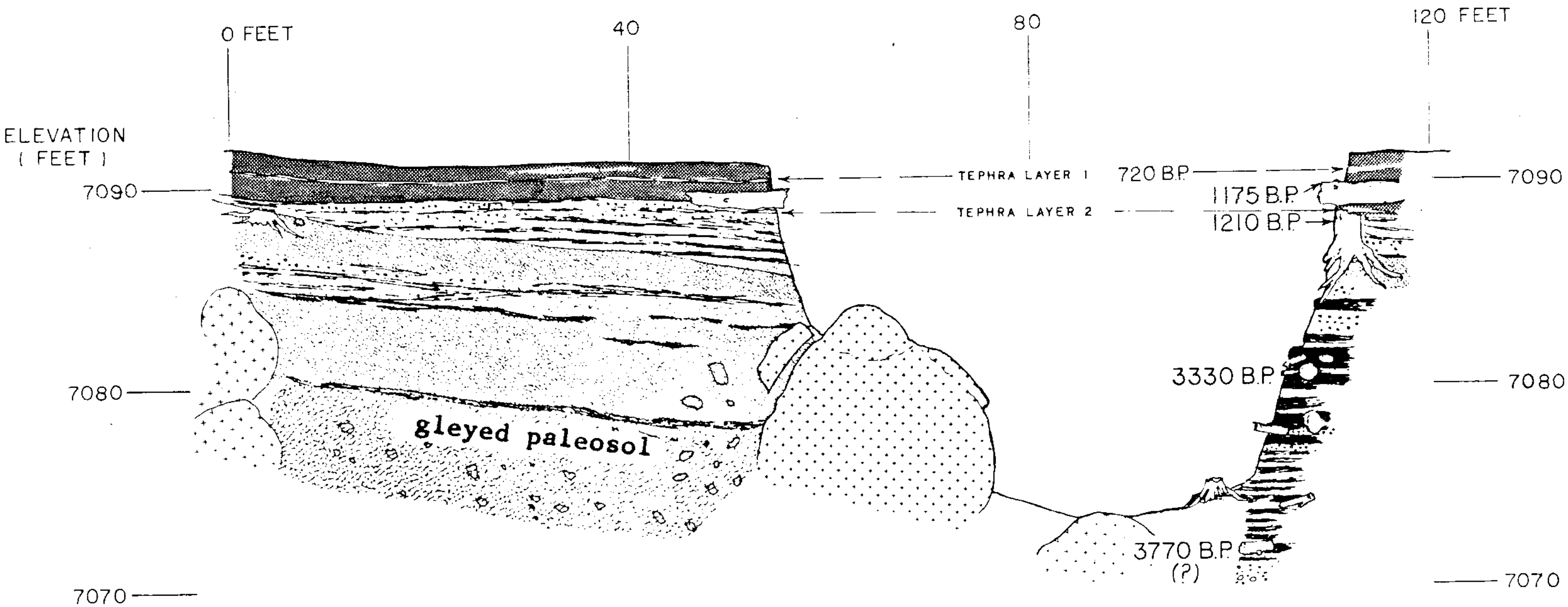


Figure 3-20. Transverse section through meadow on west fork of Long Meadow Creek, Wishon Reservoir area of Sierra National Forest, (point C on figure 3-19) showing position of gleyed paleosol and location of bracketing radiocarbon ages on Tephra 2. Large boulders are glacial erratics of the underlying till. 2 x vertical exaggeration.

suggest an interval of erosion in the middle Holocene. The sequence is similar to other meadow sections described, and the recent meadow does not owe its origin to drainage impeded by glacial deposits.

Boggy Meadow, Kings Canyon National Park

Setting

Boggy Meadow (elevation 7200 feet) lies at the foot of the steep south wall of Sugarloaf Valley, a broad upland basin just south of Kings Canyon. The major streams enter Sugarloaf Valley from the south side. Their headwaters are strung along an 11,000 to 12,000 foot altitude, glacially sculptured east-west divide. Bedrock in the Sugarloaf drainage is entirely intrusive rock.

The 8-acre meadow has an 8 per cent slope and recieves surface water from a 0.5 square mile drainage area. Small discharge and heavy vegetation have caused deposition of a 26-foot thick sequence of organic and alluvial soil at an exceptionally steep slope. The meadow location is probably determined by a bedrock surface configuration that forces seepage water in the colluvial forest soil to emerge at this point. A closed mixed-conifer forest surrounds the meadow.

A continuous gully now dissects the east half of the meadow, and a discontinuous branch is working headward along the western edge of the meadow. Remains of gully-check dams built in the 1950's occupy both branches. These may have helped save the west half of the meadow. Lowering of the water table by deep dissection in the east half allows invasion by lodgepole pine. Sharsmith (1959) reports that he was told by former district ranger, Stanley Bechtal, that the deep gullies were cut in 1926. Tree-ring age of the oldest lodgepole pine in the gully confirms that gullying began prior to 1934.

Proximity to Pleistocene glaciation

The major tributaries of Sugarloaf Valley were extensively glaciated during the Pleistocene. Impressive lateral moraines of Wisconsin age lie on the flanks of these valleys, but only the glacier in the Roaring River Valley extended into Sugarloaf Valley. A broad terrace consisting, in places, of more than 50 feet of bouldery outwash resting upon bedrock rises 80 feet above part of modern Sugarloaf Creek. The terrace was apparently constructed

Figure 3-21. Location map of Boggy Meadow and preliminary map of Wisconsin glacial deposits in valleys of Roaring River, Ferguson Creek, and East Fork of Sugarloaf Creek from Birman and Berger (1965) and reconnaissance mapping in Sugarloaf Valley. Ti = Tioga, Te = Tenaya, and Th = Tahoe. Heavy lines show prominent crests of moraines.

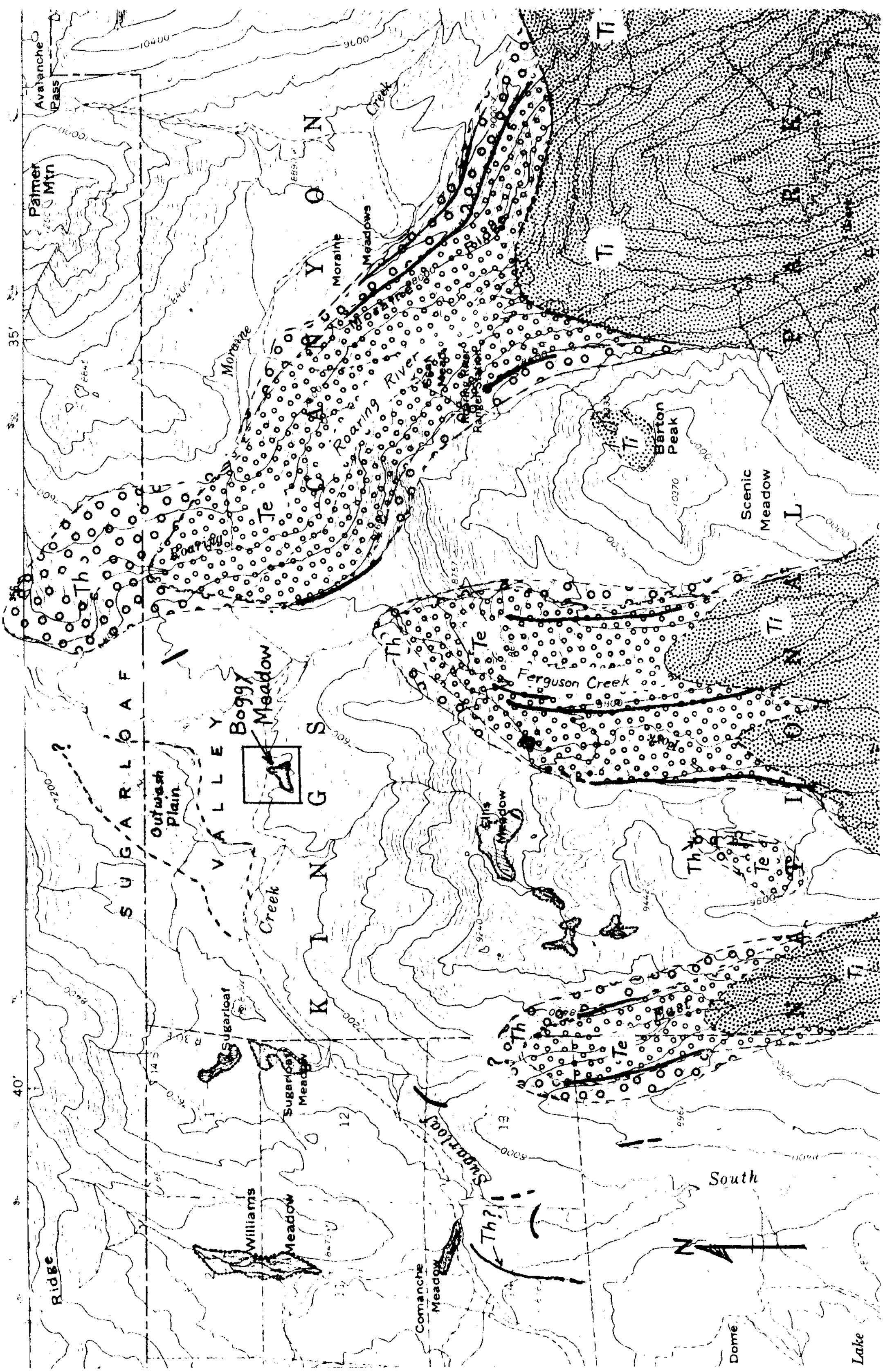

Avalanche Pass
Palmer Mtn
Moraine
Moraine Meadows
Creek
Roaring River
Barton Peak
Scenic Meadow
Ferguson Creek
Boggy Meadow
Outwash Plain
S U G A R L O A F
V A L L E Y
K I N G S
C A N Y O N
N A T I O N A L
P A R K
Creek
Ellis Meadow
Sugarloaf
Sugarloaf Meadow
Williams Meadow
Comanche Meadow
Sugarloaf
South
Ridge
Dome
Lake
Ti
Te
Th
Th?
35'
40'
0'
5000'
scale

Table 3 - 2. A tentative summary of Wisconsin glaciation in the Roaring River drainage, Kings Canyon National Park (from Birman and Berger, 1967)

advance	total volume of all glaciers (cubic miles)	down valley extent of glacier: Roaring River		down valley extent of glacier: Ferguson Creek		fresh to weathered boulder ratio
		distance (miles)	elevation (feet)	distance (miles)	elevation (feet)	
Tahoe	greater than 4.0	13.5	6400	6	7450	23:77
Tenaya	3.7	12	6640	6	7600	50:50
Tioga	3.3	8	7600	3.5	8760	71:29

as an outwash plain from material derived from glaciers in the upper forks of Sugarloaf Creek and upper Ferguson Creek at a time when the lower end of the valley was blocked by the Roaring River glacier.[3]

Distribution and relative ages of moraines in the Roaring River drainage are described by Birman and Berger (1967). They tentatively recognize Tahoe, Tenaya, and Tioga advances. The extent, and relative age criteria for the moraines of Wisconsin glaciations are summarized from Birman and Berger's work in table 3-2 and shown on a map based upon their work in figure 3-21. Neoglacial advances of Recess Peak and Matthes age are recognized in northeast-facing cirques above 9600 feet. The dissected fill at Boggy Meadow (figure 3-22) could not have been directly influenced by glaciation in the late Wisconsin, although ice or outwash may have spilled into the drainage area during earlier more extensive glaciations.

Stratigraphic section

Beneath 10 to 30 feet of stratified valley fill are sound boulders several feet in diameter that appear to be corestones weathered from the bedrock. A bedrock surface is not exposed in the gullies, but it is presumed to lie beneath the corestones. At the upper end of the dissected meadow (point S, in figure 3-22) is a deposit of sub-angular boulders and cobbles set in a silt matrix. Boulders and cobbles in this deposit have disintegrated so that they erode "flat to the face" and can be cut with a spade. The extent of weathering suggests an antiquity similar to that of pre-Wisconsin till elsewhere in the Sierra, but the local setting is such that on consideration, this could be a colluvial or creep mantle comprised solely of weathered constituents.

A point 300 ft. from the lower end of the gully, removal of slumped bank material in the bottom of the gully exposed a bed of cobbly gravel. This gravel is coextensive with inorganic gravel and sand at the base of the section in the lowermost reach of the meadow. Sub-angular granitic cobbles up to 9 inches in diameter occur in this sandy gravel. A paleosol profile, now buried under the fill, developed into this inorganic cobbly gravel bed (figure 3-23). This profile with a preserved, stratified, accretionary "O" and "A" horizon is identical in stage of development, texture, and color to the basal paleosol at East Meadow in Yosemite (figure 3-5). The charcoal layer, 0.5 cm thick at the base of the "A" horizon appeared to be the oldest datable material with stratigraphic context

footnote for page 88

[3]Identical bouldery terrace deposits lie in Illouette Valley of Yosemite National Park and are incorrectly described as lake deposits by Matthes (1930, plate 32).

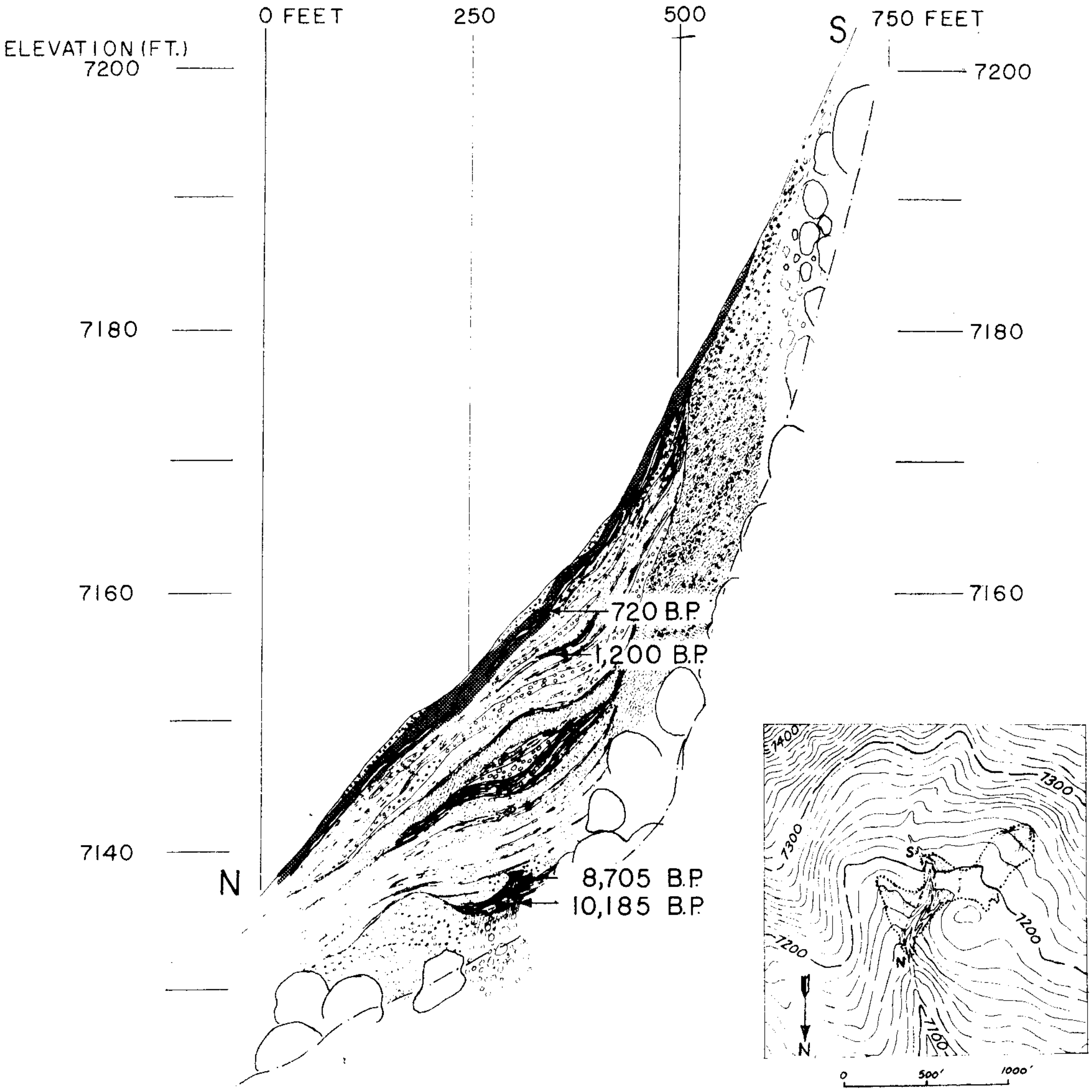
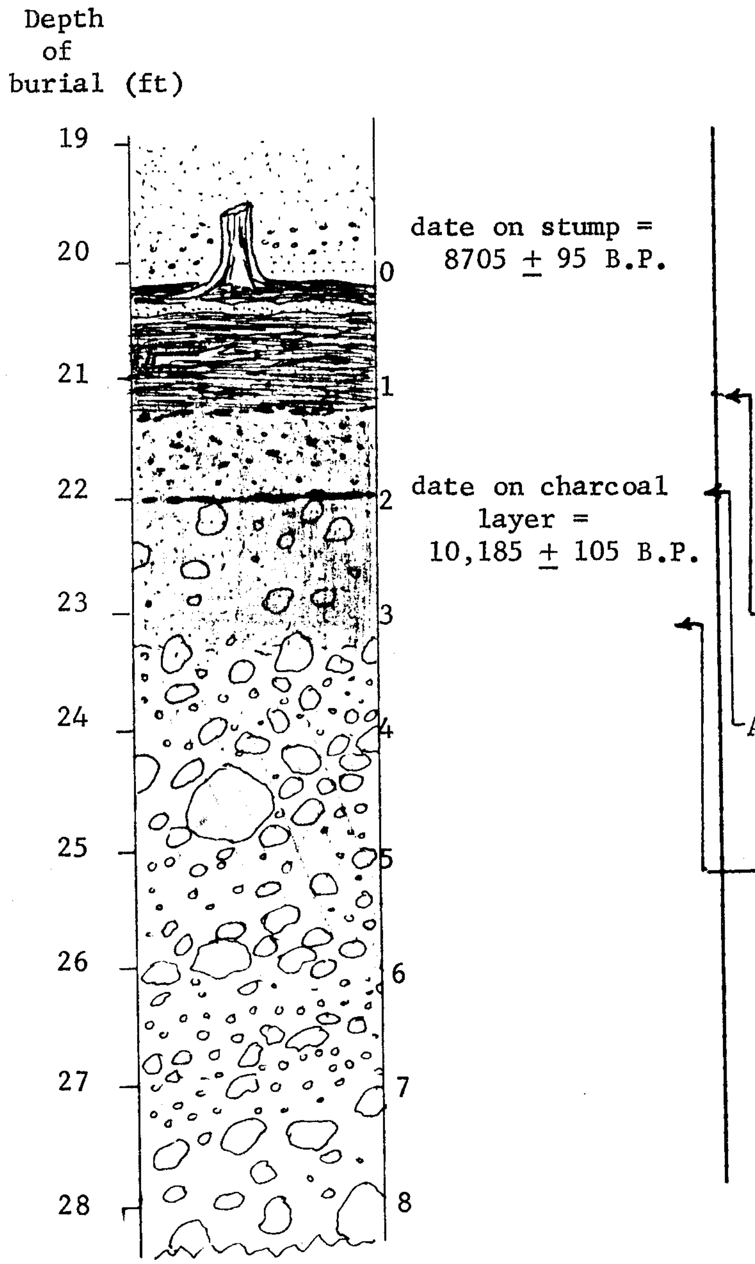


Figure 3-22. Longitudinal section through Boggy Meadow, Sugarloaf Valley, Kings Canyon National Park. Refer to legend on page 49. The 8 per cent surface slope of meadow is exaggerated 15x in order to conform with other meadow-section illustrations.



HORIZON DESCRIPTION

O : (0 to 1.2 ft), black (10YR 2/1) finely divided humus, charcoal, and silt. Consistancy sticky but friable. Tough fibrous lenses with abundant conifer needles, twigs, fine sand lenses and charcoal bands.

A : (1.2 to 1.7 ft) black, (5Y 2/2) humic sandy silt, scattered charcoal.

AC : (1.7 to 3.0) dark gray (5Y 4/1), friable silty loam with scattered cobbles.

C : (3.0 to 6.5) sandy gravel with iron-stained cobbles up to 9-inch diameter

(6.5 +) sandy gravel as above without iron stain. Some cobbles are disintegrated.

Figure 3-24. Profile description of basal paleosol, Boggy Meadow, Sugarloaf Valley, Kings Canyon National Park, California.

that could be collected from the soil. The charcoal yields a radiocarbon age of 10,185 ±105 B.P. Wood from a softwood stump rooted in the top of the "O" horizon of the paleosol yields a radiocarbon age of 8705 ±90 years B.P. This stump was subsequently buried by the stratified deposits of the next overlying unit.

The overlying 6 to 8-foot unit is composed of interstratified poorly sorted sands, sandy humic loams, woody peat, and an upper 1.5-foot thick layer of unstratified coarse sand loam soil without internal stratification. Throughout this unit, the deposits contain abundant wood fragments, roots, and charcoal. Although no stumps occur in this section, the deposits suggest intervals of forest encroachment, and generally dryer meadow environments.

Above the thick unstratified coarse layer, the deposits change to thinly stratified humus-rich sand and silt loams interstratified with well-sorted sand and fine gravel layers of transported grus indicating wet meadow conditions. This uppermost unit contains two volcanic ash layers identified as tephra 1 (720 years B.P.) and tephra 2 (1200 years B.P.) Tephra 2 occurs as discontinuous lenses less than 3 mm thick at depths of 2 to 4 feet. Tephra 1 occurs as a nearly continuous layer up to 3 cm thick at depths of 0.5 to 1.5 feet.

Chronology and interpretation

The basal cobbly gravel is pre-Holocene in age, as indicated by the radiocarbon age of 10,185 years B.P. in the overlying soil. Proximity to late Wisconsin glacial deposits suggest that the meadow site was in the relatively barren, unforested, alpine or sub-alpine belt during glaciation. Lack of organic detritus or charcoal in the basal gravels further supports the interpretation that the watershed was relatively vegetation free. No material of cobble size is found in the overlying fill suggesting that the basal alluvium was deposited under stream regimes different from any that have occurred in the Holocene. If these basal gravels relate to the last major glaciation in the Sierra, then the 10,185 ±105 years B.P. age in the soil developed upon them, presently stands as the most precise minimum age for montane forestation of the upper Sierra during the early post-glacial time. An age of 9990 ±800 years B.P. was obtained by Adam (1965) in the Lake Tahoe area for this transition.

This basal age and the 8705 years B.P. on the stump at the top of the paleosol bracket a 1480 radiocarbon year interval of soil formation represented by the paleosol. Excellent preservation of organic material in the "O" horizon indicates that the soil accumulated under wet conditions. Lack of interruption of the accretion of the silty "O" horizon by clastic deposition suggests considerable ground-surface stability provided by a thick plant cover. Charcoal in the soil indicates a forest cover and suggests that the soil developed under a wet, forested, glade environment.

The overlying material is considerably less organic than the paleosol "O" horizon and appears to represent episodic slope-wash deposition of poorly sorted clastic material upon moist to dry meadow deposits and forest soils. The un-dated woody peat near the top may represent greater soil moisture or it may represent ponding caused by local changes in seepage or surface drainage patterns. The overlying 1.5-foot thick unstratified layer is typical of conifer forest soils where good drainage allows churning and mixing of the materials by rodents, and tree fall. No chronologic control has been obtained for this middle unit, although abundant woody material is available should it become important to obtain precise ages on the various layers. The middle unit deposits represent some part of the time between 8705 and 1200 years B.P.

Above the forest soil layer ascribed to good surface drainage, are humus-rich sand and silt loams, sedge peat, and sorted layers of transported gruss typical of moist meadow deposits. Occurrence of the 1200 year B.P. tephra 2 layer in the middle of this uppermost meadow unit suggests an age greater than 2000 years for the establishment of relatively stable, open moist meadow conditions at this site.

Stratigraphy of subalpine meadows

The deposits and stratigraphy of meadows in the subalpine belt are somewhat different than the upper montane belt. Organic topsoils seen in the subalpine valleys are confined to the upper few feet of deposits. Deep gully erosion does not commonly occur in these meadows partly because a base level is formed by the underlying coarse gravel or till, and partly because subalpine watersheds are not subject to major winter floods such as have occurred in the montane belt (see Chapter 2).

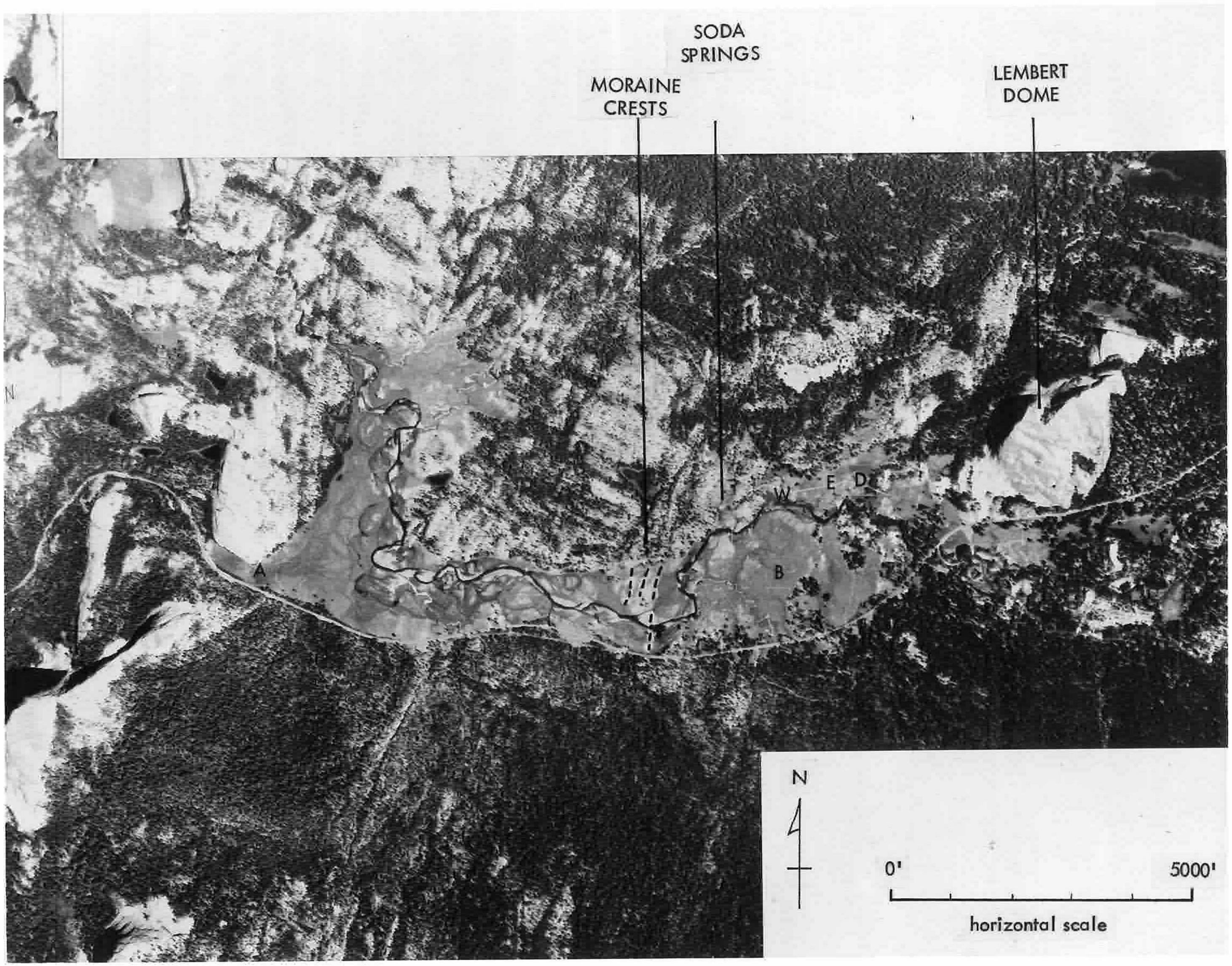
This is the belt of "glacier meadows" eloquently described by John Muir (1894). Many of Muir's ideas have withstood the tests of time and more detailed studies, but the concept of the meadow as a late stage in the successive infilling of a glacial lake is greatly oversimplified and overworked, reflecting theories of landscape evolution that dominated early 20th-century schools of geomorphology (Davis, 1909) and plant ecology (Clements, 1928). It is appropriate to examine the Tuolumne Meadows area in some detail, for the stratigraphic and chronologic data obtained there suggest that many of the subalpine meadows are not infilled glacial lakes, and that many of today's verdant meadows are relatively newcomers to the landscape.

Tuolumne Meadows, Yosemite National Park

Setting

The Tuolumne Meadows collectively cover more than one square mile and are the largest, and perhaps the most scenic of the Sierra subalpine meadows. The meadows begin near the confluence of the Lyell and Dana forks of Tuolumne River at 8600-foot elevation, and lie along both sides of a 3.5 mile reach of the river to an elevation of 8400 feet. At their widest they extend 0.5 miles across the relatively broad valley (figure 3-24). The broad valley bottom is blanketed by a sub-alpine meadow community of grasses, sedges, rushes, and herbs, surrounded on the meadow's edge by a forest of exclusively lodgepole pine. The Tuolumne River drainage above the meadows is about 90 square miles. Granitic rock underlies the meadow area and most of the Tuolumne Basin with the exception of the Dana Fork headwaters which drain several square miles of metamorphic rocks along the crest of the Sierra. Gravels in the present Tuolumne River bed are

Figure 3 - 24. Vertical air photo of Tuolumne Meadows, showing abandoned meanders of Tuolumne River and locations of stratigraphic sections and soil profiles (U. S. Geological Survey photo GS-VJS, 1-105, Aug. 8, 1955)



SODA
SPRINGS
MORAINE
CRESTS
LEMBERT
DOME
A
B
W
E
D
N
0'
5000'
horizontal scale

predominantly hornfels cobbles derived from these metamorphics. Hornfels cobbles survive weathering and fluvial transport better than granitic rocks. Differences in percentage of hornfels clasts in gravel deposits are useful in distinguishing among the bedload gravels of the Tuolumne River (50%), glacial deposits (5%±), and gravel derived locally from bedrock (<1%).

Geomorphology of the basin is discussed briefly by Matthes (1930, p. 15) and Wahrhaftig (1965, p. 123). According to Wahrhaftig, the broad, low-gradient, valley floor upon which the meadows lie (figure 3-25) would, in Matthes view, be a valley developed during the Pliocene in response to the "Mountain Valley Stage of Sierran uplift." Further uplift in the Quaternary has caused the Tuolumne to incise a deep gorge into the west slope of the Sierra. California Falls, 3 miles downstream from the meadows is interpreted as a knickpoint at the head of the Grand Canyon of the Tuolumne and represents the upstream extent of this "Canyon stage" of valley development."

The upper Tuolumne Basin was glaciated many times during the Pleistocene. There is little doubt that ice of the pre-Wisconsin stages identified on the east slope as summarized by Sharp (1972, p. 2235) covered the meadow area, but the evidence has been removed or obscured by subsequent glaciations. Abundant evidence for Wisconsin-age glaciation remains in the basin. Matthes (1930) estimates that ice was 2200 feet thick over Tuolumne Meadows during the Wisconsin Maximum when ice of the Tuolumne Field flowed westward through this area to form great trunk glacier that terminated 60 miles west of the crest of the Sierra in the Grand Canyon of the Tuolumne. Glacially polished and striated bedrock and deposits of till protrude into the meadow area. Two broad low moraines about 10 feet high lie one mile south of Lembert Dome. Features left by the last glaciation are assigned to the Tiogan stage by Wahrhaftig (1965, p. 112), although no detailed work has been reported on the relative ages of glacial deposits in this area.

Scattered protrusions of bedrock and hummocks of till throughout the meadow area suggests that the glaciers did not excavate much below the present surface. There seems no justification for the popular notion that the Tuolumne Meadows was occupied by a large lake when the last glacier first receded. Small shallow bedrock basins resulting from differential erodability of the bedrock are common in the basin, and these may have harboured small local lakes. Kettles and other irregularities in the moraine surface may have harboured ponds such as seen in the Dana Meadows area on Tioga Pass. Shallow

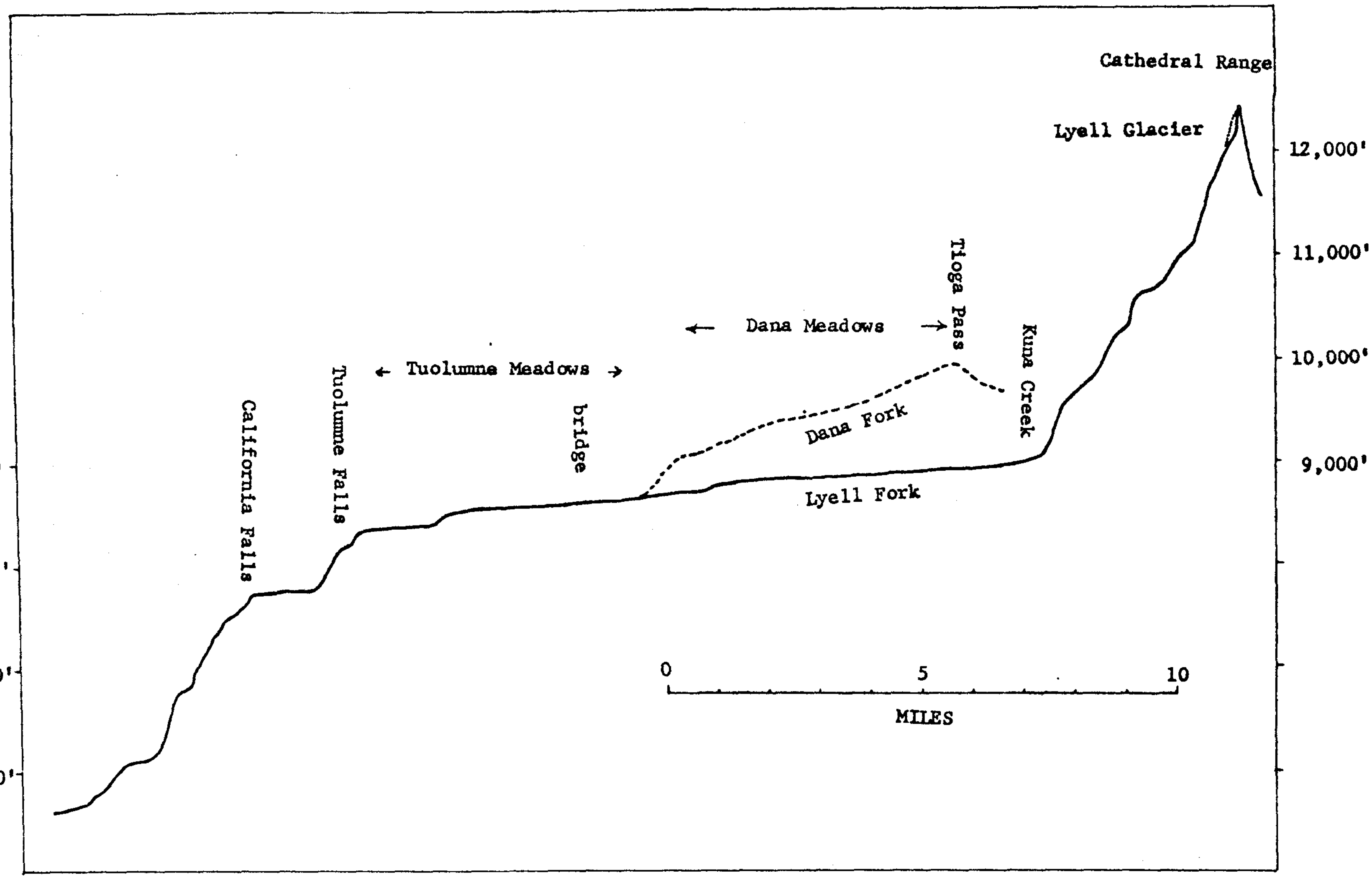


Figure 3-25. Longitudinal profile of the Tuolumne River from the crest of the Sierra to the grand canyon of the Tuolumne

coring in many such low spots has encountered only stream alluvium, however, depth of penetration was limited to a few feet, and it could not be shown conclusively in all instances that the stream alluvium rested directly upon till or bedrock.

Glacial meltwater streams, the meandering post-glacial Tuolumne River, and side tributary streams have subsequently modified much of the original surface to the present graded condition.

The stratigraphic section

In 1971, a temporary ditch exposed a four-foot section along the Soda Springs road at the north edge of the meadow providing a unique opportunity to examine sub-meadow stratigraphic relationship (fig. 3-26). Sections elsewhere in the meadow were excavated to a depth of 5 feet (sites A,B,C in fig. 3-24). Coarse gravels and a shallow water table prevented deeper investigation. These sections exhibit a similar stratigraphy. Upon the till are the river gravels or granitic sands washed in from the valley sides. Upon this alluvium rests a blanket of humus-rich sandy topsoil rarely over 3 feet thick. The topsoil clearly accumulated under meadow conditions and its base must represent the beginning of the present meadow environment for no deeper occurrences of organic soils are encountered in the sections.

Glacial deposits and soil development on the till: Glacial till exposed by the ditch, consists mostly of granitic rocks. Where the ditch cuts till hummocks, one sees an incipient soil profile. the profile, described in figure 3-27, consists of an "A" horizon, a color "B" horizon, and a "C" horizon of iron-stained, but unweathered till. The profile is two to three feet deep and is inferred to be representative of the extent of soil development on moderately well-drained granitic till in this subalpine environment since the last Wisconsin glaciation. Cobbles and gravel of the parent till occur throughout the "A" horizon though not abundantly as in the deeper horizons. The lower boundary of the "A" horizon is abrupt. Roots of meadow plants penetrate to, but not below this boundary. Pocket gopher and the Belding ground squirrel inhabit the better drained sites; and parts of the "A" horizon are churned by their activity.

The one to two foot thick "B" horizon is distinctly yellower and lighter in color and coarser in texture than the "A" horizon. Its boundary with the underlying till of the "C" horizon is gradational over several inches. The "C" horizon is unweathered till.

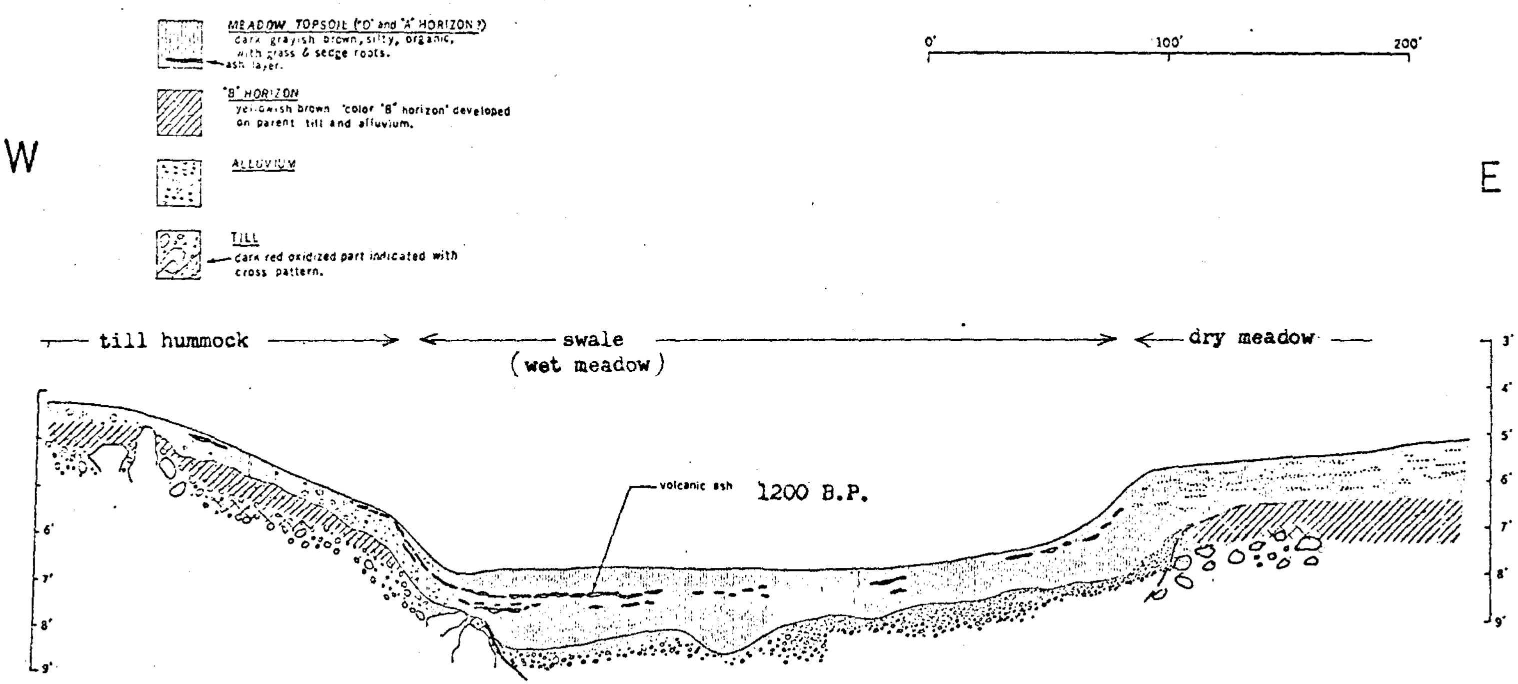
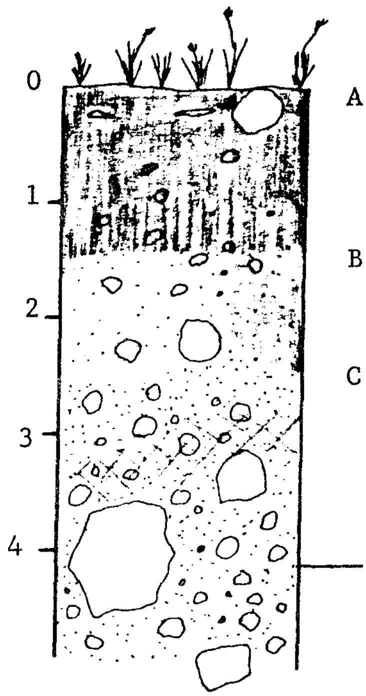


Figure 3-26. Sratigraphic section exposed in a temporary ditch along Soda Springs Road (figure 3-24), Tuolumne Meadows, Yosemite National Park. 20 x vertical exaggeration.



A: dark grayish brown (10YR 3/2) humic sandy loam. Sod of fine filamentous roots extends to 0.4 ft. Small pockets of volcanic ash 0.2 to 0.5 ft beneath surface. Scattered cobbles throughout.

B: yellowish brown(10YR 4/4) loamy sand, cobbly.

C: light olive gray(5Y 6/2) glacial till

water table (September)

Figure 3-27. Profile description of soil developed upon hummocks of late Wisconsin till, Tuolumne Meadows, Yosemite National Park, elev. 8300 ft. Slope is 1 to 6 percent, Present vegetation is Calamogrostis, short hair sedge, groundsel, bilberry, and scattered small Lodgepole Pine.

In the vicinity of Soda Springs it is commonly iron stained and cemented with iron oxides and carbonates apparently derived from the spring waters. In swales where till is overlain by accumulations of alluvium and meadow topsoil, the till is unweathered, in contrast to profile development upon the hummocks.

Fluvial gravels: In low areas, the till is overlain by subround, moderately sorted, cobbly gravels in which clasts of hornfels predominate, suggesting that they are truly fluvial and not just reworked till. These gravels in the Soda Springs ditch and at sites B and C could have resulted from meandering of the Tuolumne River as suggested by the airphoto representation (figure 3-24) of the present river and its meander swath.

Finer sandy alluvium, in which hornfels constitute less than 5% of the clasts, rests upon the fluvial gravels and at places directly on till. These deposits are less than 2 feet thick and are regarded as the product of side tributaries because they lie within possible courses of side washes and streams. At site B these sandy fine gravels contain 5 layers rich in charcoal (figure 3-28). A blanket of meadow topsoil has accumulated on top of these gravels so their age is critical to dating the initiation of meadow conditions. It was possible to collect a sufficient amount of charcoal from one of these layers for an age determination. This charcoal yields a date of 3930 years B.P. From stratigraphic relationships throughout the meadow, this appears to predate all occurrences of organic rich meadow topsoil in both the Tuolumne and the Dana meadows.

Meadow topsoil: Overlying the alluvium, and depending upon topographic position, the till, is a 1 to 3 foot layer of topsoil, best described as a humic sand or silt loam. Detailed description of this topsoil is given in figure 3-28. Throughout the meadow, contact between the topsoil and the underlying gravel is sharp. The lowest third of the layer is usually a little lighter and yellower in color than the topsoil closer to the surface, but the layer has an overall uniform dark-brown color. The lower third is often mottled and characteristically has a higher pH value suggesting slight oxidation of the iron and soil organic matter by the seasonally fluctuating ground water.

The topsoil has accumulated to a maximum thickness of 3 feet in the low areas. Here it is stratified as shown by thin sandy lenses and layers rich in fine charcoal. In the middle of the topsoil layer is a conspicuous, but discontinuous, layer of white pumiceous volcanic ash and fine lapilli identified as 1200-year-B.P. tephra 2. At the better drained sites, the ash occurs only as pockets or is disseminated over the upper few inches,

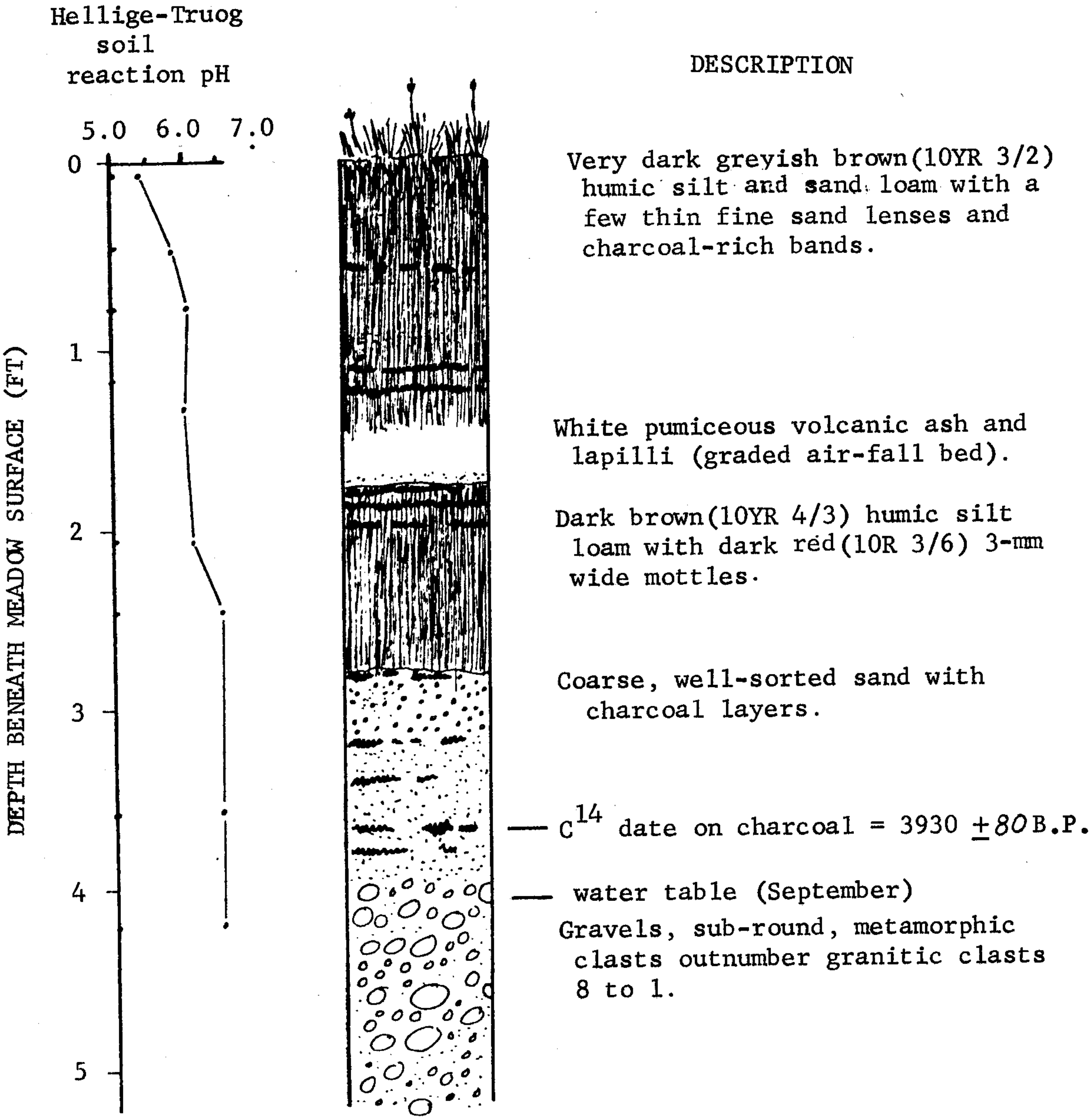


Figure 3-28. Description and stratigraphy of meadow topsoil at site B (figure 3-24), Tuolumne Meadows, Yosemite National Park, California.

giving the topsoil a lighter color. In the swales it reaches a thickness greater than 3 inches, and the base of the ash layer shows graded bedding with fine lapilli up to 0.5 cm in diameter at the bottom, overlain by finer ash typical of an air-fall origin. This ash is widespread in the northern Yosemite area and invariably occurs in the middle of the 1 to 3 foot layer of meadow topsoil. At locality B and at station 1035 in the Soda Springs Road ditch a second thin ash layer is found a few inches deeper in the soil. This could be correlative with the deeper, whispy, ash layer at East Meadow whose age was estimated to be between 1400 and 1670 years B.P. by peat accumulation rates.

The occurrence of the 1200 year B.P. tephra 2 in Tuolumne Meadows shows that the topsoil both in the lows and also on the flanks of meadow covered hummocks has accumulated by contributions from airborne and waterwashed fine material and the decomposition products of plants. Where the entire accumulation is preserved, as in the low areas, the rate of deposition of humic silt loam is between 1 and 1.5 feet per 1200 years or about 2.3 cm per century. Extrapolating with this rate to the base of the topsoil layer gives an estimated age ranging from 2000 to 3600 years B.P., but most sections indicate an age of about 2300 years B.P. This is an astonishingly young age for the establishment of meadow conditions in the Tuolumne area, but it agrees with the estimated age of 2500 years B.P. for the establishment of East Meadow of Aspen Valley.

Interpretation

The stratigraphic sequence in Tuolumne and many other sub-alpine meadows is till, fluvial gravels, and meadow topsoil containing late Holocene Tephra 2. Lacustrine silts of glacial meltwater origin are rarely encountered. Rocks and till protrude the meadow surface and total post-glacial deposits in most subalpine valleys are thin, probably less than 10 feet. Present rate of sediment production in the subalpine belt is small (Jahnda, 1967). This leads to the conclusion that Tuolumne Meadow was not a large successively infilled lake. Small depressions left by receding glaciers were probably infilled during de-glaciation or shortly thereafter by the Tuolumne River reworking glacial debris and laying down well-sorted gravels. Lack of meadow topsoils older than 2300 years B.P. can be explained in two ways. If meandering streams are in a continual process of eroding topsoil down to the gravel base level, then only recently deposited topsoils will be encountered. But this process does not explain the relatively uniform age of the base of the topsoil over the entire meadow. The preferred explanation is that changing hydrologic conditions associated with the onset of neoglacial climates in the Sierra caused a rise in the valley-bottom water table. Without a shallow water table, well-sorted alluvial gravels are not easily colonized by meadow plants for these gravels have little water retention and are unable to draw from a deeper water table by capillarity.

Discussion

Generalized depositional sequence

Preceding descriptions of stratigraphy exposed in dissected meadows within that part of the west-slope Sierra montane belt between 5000 and 7500 feet indicate a generally comparable depositional sequence beneath meadows. The complete sequence is observed in small drainage basins, floored by intrusive rocks, of the broad unglaciated southern Sierra uplands. The degree of similarity to the generalized sequence of deposits beneath meadows in major glaciated river valleys and in basins of the northern Sierra floored by volcanic rocks is not known. The generalized sequence summarized in figure 3-29 is described starting at the base:

1) A basal layer several feet thick of coarse, inorganic alluvium containing cobbles and small boulders, and resting upon an irregular bedrock surface or upon corestones derived from the partial disintegration of granitic bedrock. In all cases this basal unit appears to be pre-Holocene, and it is thought to consist of stream deposits formed under a hydraulic regime supported by the latest Wisconsin full-glacial conditions.

2) A buried soil profile extending 3 feet into underlying alluvium or bedrock and gleyed at some localities. The well-preserved, accreted "O" horizon yields radiocarbon ages between 8705 and 10,185 years B.P. Stumps of montane-belt forest trees occur in this paleosol. During this interval of soil formation, muck and fine sand accumulated in swales.

3) A 10 to 20-foot unit of stratified sandy deposits. These deposits exhibit no profile development. They are composed of coarse sand-loams with charcoal, logs, and stumps of fir and yellow and lodgepole pine. Interstratified, clastic, layers are poorly sorted. This sequence reflects episodes of good soil drainage and forestation of the valley bottom. Its age is not precisely bracketed but is known to lie between 8700 years B.P. and 1200 years B.P., with 2500 years B.P. being the younger limit at some localities. Radiocarbon ages of 2830 and 3320 and 3770 are obtained from within the unit. Limited chronologic control suggests scour and fill complicate this unit, and up to 10 feet of material may have been deposited within a few hundred years.

4) A 2 to 12-foot thick unit characterized by interstratified layers of sedge peat,

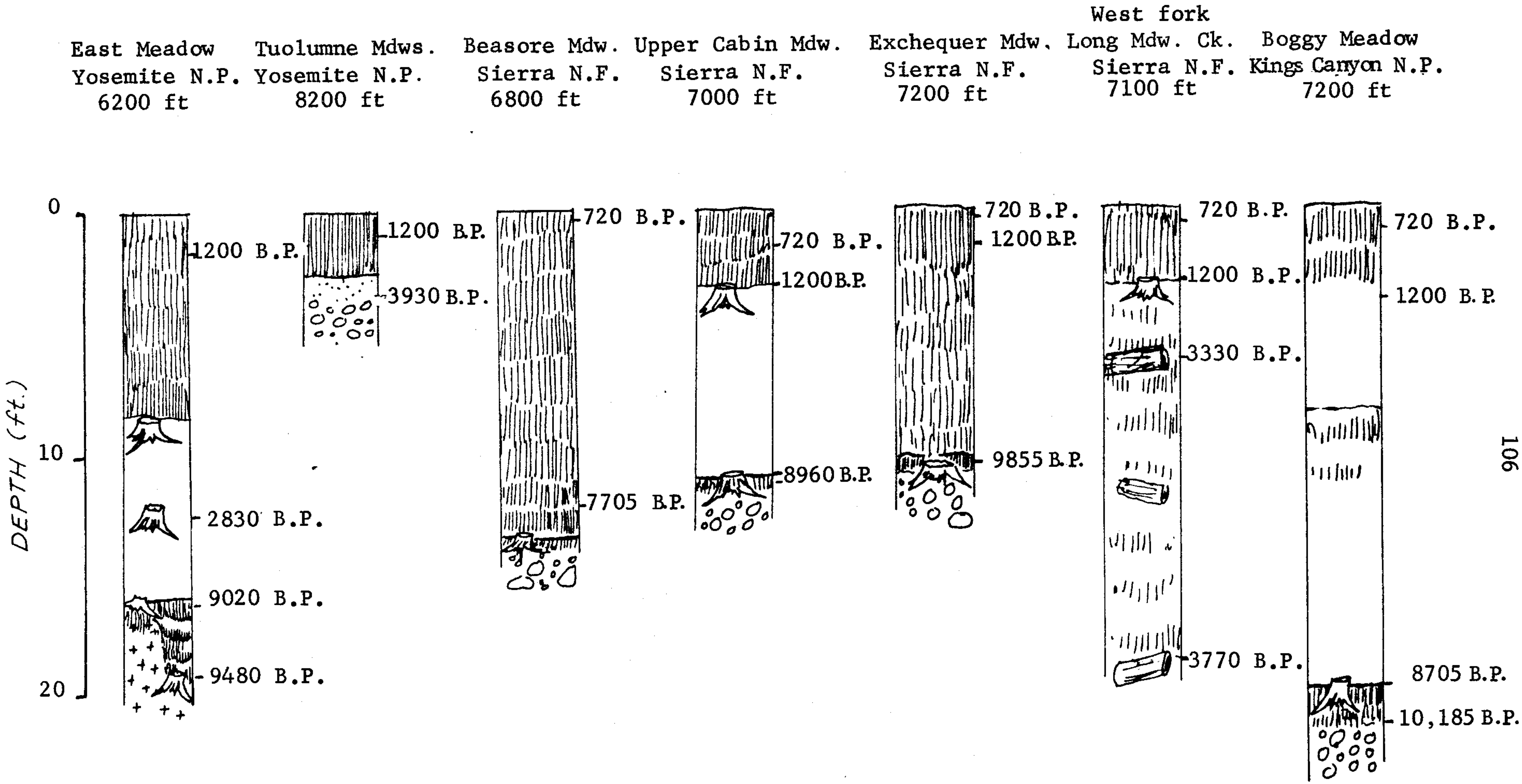


Figure 3-29. Summary of stratigraphy and chronology beneath west-slope meadows of the southern Sierra Nevada.

humus-rich sandy loams, and layers of sorted transported grus . This unit is similar to material presently being deposited on the surface of wet meadows. It represents the establishment of open, wet meadow conditions. Ages within the unit are 1545 B.P., 1200 B.P., and 720 B.P. The transition to this wet-meadow condition is estimated to lie between 2500 and 2300 years B.P. at some sites, and is well-dated at 1200 years B.P. at two sites.

Causes of litho-stratigraphic change in layers

Distinguishing characteristics of depositional units in the valley fill involve: organic-matter content, fossil evidence of the plant community, texture, internal stratification, and in one instance, soil-profile development. Although these characteristics have changed during Holocene aggradation they remain relatively consistent within depositional units for intervals that lasted one to several millenia and then abruptly changed. Nature of depositional layers on the valley bottom is a result of the hydrologic processes and the plant environment prevailing at the time. Chemical pedogenic processes and mineral alteration appear to be important only in the formation of the basal paleosol.

Striking contrasts in the appearances of layers result from widely ranging organic-matter content and texture differences involving the clastic component. Organic-matter content of soil reflects the climate-related variables of soil-moisture content and temperature. Both variables control the rates of biomass production and organic-matter destruction by organisms or oxidation. The relationships of these variables to these rates are known only in a general way, but Mohr and van Baren (1954) provide a qualitative assessment of the affect of these variables on humus accumulation in tropical soils (figure 3-30). Their temperature scale and relative rates cannot be directly applied to the Sierra montane belt, but figure3-30 does illustrate that a change to good drainage or an increase in summer temperatures may reduce the organic-matter content of topsoil.

The amount and grain size of the clastic component reflects the nature of sediment-transporting flows on the valley side slopes and across the depositional site which are in turn affected by topography, plant cover, and surface water hydrology discussed in Chapter 2.

Most of the deposits can be interpreted through identification of modern counterparts

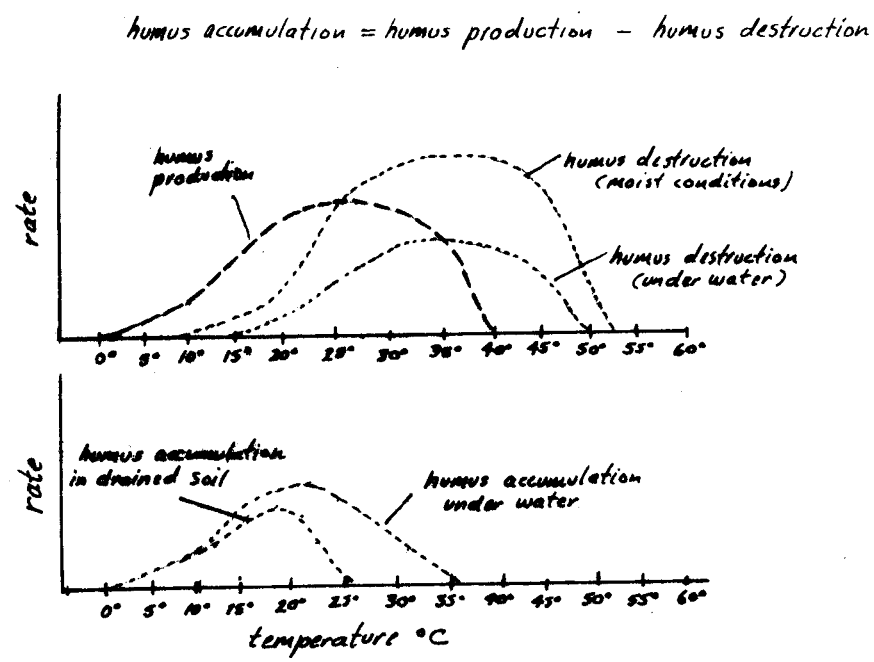


Figure 3-30 . Relative rates of production,destruction, and accumulation of humus in tropical soils based on consideration of relationships of soil temperature and moisture conditions to rates of plant growth and activity of soil microorganisms (from Mohr and van Baren, 1954).

of the buried layers. Great difference in organic-matter content and in the texture of the clastic component are found in topsoils of the two principal plant communities: the closed conifer forest and the open wet meadow. Existence of the two communities in valley bottoms is determined solely by depth to water table. Growing season water tables shallower than 2 feet support a meadow and inhibit growth of coniferous trees. Conifers in the Sierra appear to grow to maturity where water tables are deeper than 5 feet.

Depth to the water table is here treated as a quantitative measure of soil drainage which is taken to be the natural subsurface removal of excess-near surface water. Excess water is that which maintains the near-surface soil moisture content at field capacity (saturation). Soil drainage expressed as the configuration of a shallow water table in valley fill is a function of four variables: 1) geometry and thickness of permeable fill, 2) permeability of the fill, 3) supply of subsurface water (including local infiltrated surface water), and 4) the evapotranspiration rate of the plant community. A steady state relationship is formulated (Appendix B). The situation of valley bottom sites is normally one of a shallow water table. Changes in water-table in response to the variables listed have apparently caused the water table depth at some sites to fluctuate across the critical depths required for meadow and forest plant communities. The following considerations show how changes in water-table depth may have come about.

1) As the valley aggrades, the increased thickness of fill above the water table effectively increases depth to water table.

2) The fill may become entrenched below the water table level causing its lowering.

3) Sheet erosion of fill effectively raises the water table by lowering the ground surface.

4) The plant community is reduced by fire, insect attack, or logging causing a diminished evapotranspiration rate which results in a rise of water table.

5) The upslope supply of infiltrated water changes because of climatic variation or alteration of vegetation or soil.

6) Surface drainage at the site is dammed by beaver workings, landslides, log jams, or tree falls, causing a rise in ground-water level.

These possibilities will be examined in terms of damming, erosion and deposition of fill, forest fires, and climatological change.

Damming of the surface drainage

Beaver currently reside along the east edge of the San Joaquin Valley and many attempts have been made to introduce them into the upper Sierra with little success in establishing permanent colonies (Storer and Usinger, 1968, pages 39 and 340). If beaver had naturally occupied these valleys, one would expect abundant knawed logs and other evidence in the stratigraphic section as reported by McCulloch and Hopkins (1966) in Alaska. No such material is found in well exposed sections.

A buried log jam in the west fork of Long Meadow Creek contains a log dated 3320 years B.P.; however, the slope shown in a longitudinal section through the meadow (figure 3-19) indicates that the log jam could not have provided a dam high enough to cause a change from forest to meadow . Log jams were not found in any of the other meadow sections.

No evidence is seen of former landslides or rock falls blocking the drainage of meadow areas. The possibility of overlooking evidences of damming are slight because greater-than-2-per-cent slope in meadows requires a single dam, or a series of dams with a total height of 20 to 40 feet.

Erosion and deposition of valley fill

Stratigraphic sections examined in this study do not exhibit infilled gullies leading to a field conclusion that aggradation has been a continuous process. However radiocarbon dating suggests possible hiatuses and short episodes of rapid deposition of material (10 feet within 400 years) within the interval c.a. 8700-2500 years B.P. associated with forest cover and colluvial material. Therefore, cut and fill may have been important in the middle Holocene history of valley-fill deposits. Forest floors frequently cleared of understory litter and vegetation by fire might be susceptible to erosion and provide an abundant sediment supply.

Present meadow surfaces appear to be aggrading. Meadow sod protects the surface from erosion. For erosion to occur, sod must be damaged either by trampelling, overgrazing, or by a significant increase in depth to the water table. Meadow protected surfaces commonly build up to slopes of 2 - 4 percent depending upon drainage basin size (figure 2-1). At some inclination, the surface is marginally stable, and a surface disruption presumably initiates erosion. Rate and pattern of gully dissection of modern meadows suggests that once gully erosion begins, it leads to nearly complete removal of valley fill. Conifer trees invade high standing terraces flanking gullies, but combined processes of continual ground water sapping and episodic stream erosion consume these terraces.

These modern examples of meadow dissection result in a boggy meadow surface near the bedrock base level of erosion. Presumably the aggradation process starts anew. If dissection of this type is naturally initiated there should be a few instances of pre-Twentieth-Century gullied meadow surfaces. No such examples are known. Dissection of all meadows listed in Appendix A appears to have initiated since 1900 A.D.

Aggrading meadow surfaces do not appear to build up above the water table. Fine meadow topsoil is sufficiently impermeable so that it effectively seals off seepage outflow areas and the water table apparently rises with the land surface.

Deposits of permeable sand, less than a few feet thick, build a surface above the water table, but these are eventually colonized by meadow plants and sealed off by topsoil.

In conclusion, there are no known instances in which erosion or deposition leads to major drainage changes in the valley fill without removing all of the fill. It is hypothesized that minor cut and fill may account for the variability of layers associated with forest cover that lie between the basal paleosol and the meadow deposits.

Forest fires

Effects of fire on the watershed hydrology in the Sierra are known only in a general way. Little quantitative data are available. Removal of trees by fire should significantly reduce evapotranspiration losses resulting from interception and transpiration. Transpiration losses by conifers are a significant portion of the total water budget. Summer soil-moisture depletion by a mature conifer forest may amount to 22.6 inches of a total precipitation of 68 inches at a 3000 foot elevation montane site in the northern Sierra (Ziemer, 1968).

Fire is also known to increase water repellancy of soil, thus increasing surface runoff; however, by decresing infiltration, this effect would be much less if much of the precipitation occurs as snow. Major fires in the watershed should lead to erosion of barren soil from the valley slopes and deposition of charcoal-rich deposits in the valley bottom.

Judging from the effects of clearcutting of forest trees, fire removal of the vegetation reduces the interception and evapotranspiration losses and creates openings allowing sunlight to fall on previously shaded ground where the meadow community would not grow. On sites with marginally good drainage for forest growth, clear cutting has brought about water table rises of about 1 foot (Wilde and others, 1953, Heikurainen, 1966, and Heikurainen and Päivänen, 1970). The calculations presented in appendix B using parameters appropriate for sloping valley-fill deposits of Sierra meadow sites show that removal of a forest transpiration draft on the shallow ground water also causes a water table rise of about 1 foot, but not the several feet necessary to cause a forest meadow transition.

The recurrence interval for fires at any given spot in the west-slope mixed conifer forest is about every 8 or 9 years (Kilgore, 1973, Wagener, 1961). These are not major fires but small spot fires confined to a few tens of acres. The longevity of a natural stand in mixed conifer Sierra forests is several hundred years, excepting the ancient and fire resistant Giant Sequoia. This may be an indication of the recurrence interval of holocausts. Five charcoal layers in deposits younger than 1200 years at East Meadow (pages 59-61) support the view that major fires in this watershed recur at intervals of 250 to 300 years. Uncontrollable historic wild fires that have occurred on the west slope probably covered an area comparable to major fires in the past. These fires are apparently limited more by topographic divides and weather changes than by fire fighting efforts. The largest historic fires have not exceeded 100 square miles.

Studies of fire-scarred trees (Waganer, 1961, Kilgore and Taylor, 1972) suggest "good ignition years" in 1795 and 1843 A.D. These years probably represent summers of extremely low humidity during which many large fires occurred. It would seem that extremes in annual weather could account for synchroniety of hydrologic changes in widely separated watersheds, but the recurrence interval of extremes in fire weather and the natural longevity of stands seems to be less than several hundred years, and too repetitious to account for the dated hydrologic changes reflected in meadow stratigraphy.

Major fires in the watershed should leave a clear imprint on the stratigraphic record in the form of charred stumps and logs, abundant charcoal, and burned layers. Records of past fires are abundant in the sections, but fire horizons do not seem to correlate with major changes in the character of the deposits. Chronologic control on changes in character of these deposits indicated episodes measured in millenia, rather than the few-centuries periodicity of fires. The nature of some stratigraphic sequences requires water table rises of several feet which seems too great for any fire. Although fire may be an important triggering mechanism to bring about change in the vegetation and deposits, chronologic sequence presented in this study indicates that hydrologic stress on the environment is initiated by other secular changes spaced in time by several millenia.

Climatic change

Clearly, climatic change from the late Wisconsin full-glacial conditions to the Holocene post-glacial conditions should have greatly affected the hydrology of these mid-elevation Sierra watersheds.

Secular climatic variation during the Holocene has not been so dramatic, but subtle changes have produced 300 km shifts in the arctic tree-line (Sorenson et al, 1971) as well as numerous changes in vegetation patterns inferred from pollen diagrams in North America and Europe. Alpine glaciers have been regenerated and wasted away several times during the Holocene (Porter, 1974, Denton and Porter, 1970, Benedict, 1973, and Curry, 1968). Holocene climatic variation may have had a controlling effect on the survival of people in regions where the resources are sensitive to climatic variation (Lamb, 1966, Schneider and Dickinson, 1974, and Bryson and Ross, 1975). Understanding climatic variability has taken on new urgency in our times (Schneider, 1974), for as population puts pressure on land and water resources throughout the world, minor climatic variation of the magnitude experienced in the Holocene could spell disaster to certain regions.

Although the agents of climatic change probably operate on a global scale, different regions of the earth are known to respond in contrasting ways. For instance, cool, wet periods in the high latitudes are accompanied by dry periods in the semi-tropical monsoon lands (Bryson, 1974). Therefore it is appropriate to consider in regard to the Sierra only the northern temperate climates of the west coast of North America which are largely determined by air masses moving inland from the Pacific Ocean. Climate of the Great Basin, the interior of North America, the eastern seaboard, or elsewhere in the world may respond very differently.

Details of Holocene climatic change pertinent to the Sierra are emerging from studies of pollen stratigraphy (Adam, 1965), lichenometric dating of glacial moraines and an analysis of historical records (Curry, 1968) and tree-ring climatology (LaMarche, 1974, and LaMarche et al, 1974). Tree rings in the bristlecone pine appear to provide a sensitive and precise paleoclimatic indicator that offers hope of obtaining quantitative information on temperature and percipitation variations back to about 5500 years ago (LaMarche, 1974). Datable glacial moraines in the Sierra of Holocene age appear back to about 2600 years old. No glacial advances clearly of Holocene age have been identified prior to that time. A complete record of Holocene pollen fluctuations has been studied by Adam (1965) in the vicinity of Lake Tahoe. Although certain species of plants are clearly restricted in altitudinal range in the Sierra, it is difficult, if not impossible, to identify pollen at the species level. Furthermore, members of a particular genus may be found at all vegetation belts as is the case with Pines, and the narrow vegetational belts of mountain ranges allows wind transport and mixing of pollen from various altitudinal belts. However, integrated study of the soil, sediment, and the fossil plant and pollen record at a site may allow an environmental reconstruction of the local habitat which often has paleoclimatic significance. At present, pollen records have provided only a qualitative assessment of climatic change in the Sierra. Nevertheless stratigraphic accumulations and their pollen record can be dated by radiocarbon, and they have provided the only available chronology of early Holocene climatic changes in the Sierra.

There is broad agreement between the magnitude, sense, and timing of changes in the interpreted valley-fill drainage history in this study and Holocene climate changes inferred from studies of glacial moraines and pollen. Consideration of other possible causes for changes in the nature of deposits beneath meadows suggests that such causes are not of sufficient magnitude, or that they have an inconsistent timing. In view of the ineffectiveness of these other processes, I conclude that secular climatic variation has been the overriding influence in producing the major variations in deposits preserved in the meadow stratigraphic sections.

The purpose of the following discussion is to outline what is known of the Holocene climatic record in the region and to infer the character of the climate from the nature of the layers.

Late Quaternary Climate in the Sierra

Late Wisconsin climates

The virtual lack of charcoal and other visible organic materials and the presence of large cobbles and boulders, up to 3 feet, support the view that the basal materials were deposited by flood flows of greater competency than any experienced in the Holocene, and that the watershed was sparsely vegetated. Absence of still older Pleistocene fill in the valleys indicate that prior to the Holocene, these valleys were cleaned out.

From consideration of modern glacier budgets in the Sierra, Curry (1968) estimated that a 2.5 fold increase in the mean annual snowfall (1930-60) could produce full glacial conditions, and that if summers were substantially cooler and cloudier than at present, a 1.3-fold increase would be sufficient to account for late Wisconsin glaciation. In either instance, the effect on watersheds below the firn line would be higher annual peak stream flows from snowmelt, and greater streamflow throughout the year.

Climatic change also indirectly affects runoff and peak flows by altering the vegetation (Schumm, 1965). It is likely that the meadow watersheds were above timberline during the late Wisconsin and lack of vegetative cover alone could account for higher peak flood flows resulting in greater stream competence. Consideration of estimates of the modern and Wisconsin (?) climatic firn line (Table 3-3) suggests that during full glacial conditions timberline in the Sierras may have been depressed about 3200 feet. This is consistent with Wright's (1971) conclusion from palynological evidence that during Wisconsin glaciation, tree line in the Western Cordillera was lowered 2600 to 3300 feet. A lowering of 3200 feet would place timberline 7000 to 8000 feet, near the elevation of the

Table 3-3

Estimated depression of vegetation belts in the west-slope Sierra (37° N) during the Wisconsin (?) obtained by applying modern elevation differences to estimates of the Wisconsin firn limits

	modern	*Wisconsin (?)*
Climatic firn limit	14,000[1]	10,000[2]
Orographic firn limit	13,200	9,000[2]
Timberline	11,000	6,800
Upper limit of montane forest	8,500	4,300

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[1] Flint (1957)

[2] Wahrhaftig and Birman (1965)

watersheds in which the upper montane meadows are presently situated. Climatic conditions at timberline altitudes are reviewed by Daubenmire (1954), Wardle (1965), and LaMarche (1973). A good correlation exists between timberline altitude and the level at which the mean temperature of the warmest month does not fall below 10° C. The northern limit of trees (the artic tree line) also coincides approximately with the 10° C isotherm of the warmest summer month in both North America and Asia (Löve, 1970). If the basal gravels are of late Wisconsin age the lack of charcoal might mean that the mean monthly summertime temperature did not exceed 10° C (50° F) at the meadow sites (6500 to 7000 feet) between latitudes 37° and 38° N on the west slope of the Sierra.

Pluvial events correlative with increased runoff during glaciations in the Sierra have been identified in the San Joaquin Valley west of the Sierra, and in the Great Basin to the east. Radiocarbon ages are obtained from materials associated with lacustrine sediments or with lake levels. In the San Joaquin Valley, the Kings, Kaweah, and Tule rivers discharge into Tulare Lake basin. At the present time the Kern River discharges into Buena-Vista Lake basin, but in the past it may have also flowed into the Tulare basin. Two lacustrine units were identified in the upper 70 feet of section during the drilling of test holes to study the water bearing deposits (Croft, 1968). Wood fragments of the giant Sequoia (Schmidt, 1972) from just beneath the lower clay unit yield a radiocarbon age of 26,800 ±600 years B.P. (Croft, 1968). Wood from sand layers between the lacustrine clay units yield ages of 13,350, 14,060, and 17,130 years B.P. These younger ages are interpreted as an interval of dessication. An age of 9040 ±300 years B.P. on Sequoia wood within an upper clay unit (Croft, 1968, Schmidt, 1972), dated only in the Buena Vista Lake beds, may owe its origin to a possible diversion of the Kern River from the Tulare basin to its present sink in the Buena Vista basin (Croft, 1968). Schmidt (1972) reports that four more wells have encountered buried Sequoia wood in the southern San Joaquin Valley indicating that more information on the chronology of this closed basin system may be available to establish relationships with the hydrologic events of the Sierra.

Most radiocarbon ages bearing on late Wisconsin pluvial events in the Great Basin are subject to controversy (Broeker and Kaufman, 1965, and Morrison and Frye, 1965) because of interpretation of, and geo-chemical corrections applied to, carbonaceous materials of lacustrine origin. More reliable dates of 10,500 and 11,000 years B.P. are obtained from terrestially photosynthesized material (land plant remains in rat middens) from Kramer Cave, a site 250 feet below the highest Wisconsin lake level of the Lahontan system (Broeker and Kaufman, 1965). Using these data and computed corrections for the lacustrine materials, Broeker and Kaufman (1965) concluded that the last major pluvial extended from c.a. 8,000 to 22,000 years B.P., interrupted by a short period of dessication encompassing the dated terrestial material from Kramer Cave. Morrison and Frye (1965) note that rigid application of the computed corrections leads to serious discrepancies in the stratigraphic relationships of the Lahontan system, and that many of the lacustrine materials yield ages 6000 to 8000 years too young. In view of these conflicts, one can only conclude that the last major pluvial included the time traditionally referred to as late Wisconsin and that it is older than 8000 years B.P.

Early post-glacial climate

Bracketing dates on paleosol formation define a specific early Holocene climatic period in the Sierra between 10,185 ±105 and 8,705 ±90 years B.P. The most striking feature of this interval is that its paleosol formation terminated near synchronously at widely separated sites on the west slope. Such a climatic episode was originally implied in Adam's (1965) definition of pollen zone 3 at a site near Lake Tahoe. The 9900 ±800 B.P. age for this zone was obtained from a 10-cm cored interval of finely divided micaceous peat 50 cm above the first appearance of significant amounts of organic material and montane-belt pollen. The top of zone 3 (Adam, 1965) is less clearly defined by a decline in pollen proportions of sage (*Artemesia*), TCT (*Libocedrus, Juniperus,* or *Sequoiadendron*) and fir (*Abies*) which Adam infers to be indicative of warmer temperatures as this early post-glacial climatic interval ended.

Interpretation of the paleosol is difficult because no analogues are identified in modern soils of the Sierra. Paleosol profiles developed on bedrock at East Meadow and on sandy gravel at Boggy Meadow resemble montane forest soil except for their organic-rich silty "O" horizon. The paleosol at Cabin, Exchequer and Beasores meadows have similar "O" horizons but, instead of a forest soil profile, a strong gley horizon formed in the underlying permeable gravel. In order to understand conditions at the time gley horizons formed, we must discern whether the gley is a feature of paleosol formation or the result of burial and inundation by water during the subsequent meadow history. Bunting (1967, page

125) states "a gleyed soil is one with part or most of its profile waterlogged, therefore undergoing reductions instead of oxidation. Ferric oxides are reduced to ferrous salts giving the soil a uniform grey or blue color." The gleying process in which reduced iron is translocated or formed in lower horizon is not entirely understood.

Experimental work by Bloomfield (1951) shows that fermentation products of grass cause rapid reduction of ferric ion in solution, but that peat and raw humus are essentially inactive. On the basis of Bloomfield's experimental results, I conclude that the gleying process requires a cover of vegetation not far above the gleyed horizon. The subsequent vegetation surfaces, represented by the 10 to 15 feet of interstratified meadow and forest deposits, soon became too far removed from the gley horizon in the paleosol to have had anything to do with its formation or alteration. The overlying fill at Cabin, Beasore, Exchequer, and the West Fork of Long Creek meadows, has simply preserved the gleyed character of the paleosol profile and has had nothing to do with forming it.

Formation of gley horizons requires stagnant ground water rich in fermentation products from decaying plants. It is difficult to account for stagnation of ground water in permeable gravels. One explanation is that during part of this soil forming interval the subsoil remained frozen during most of the summer growing season. Such conditions produce gley horizons in the Alps (Bouma and van der Plas, 1971). Cool summers could explain both accumulation of organic-rich "O" horizons and seasonally or perennially frozen subsoil. Alternatively, the ground water may have simply stagnated because of the low gradient ($<1\%$) of the valley bottom.

Remains of montane belt vegetation in this soil (Appendix D) tell us that the maximum mean monthly summer temperature was above 15^{o} C, but probably did not exceed 25^{o} C, and precipitation was generally above 35 inches per year.

Nature of the paleosol allows intervals of cooler and moister climate than presently occur in the Sierra, but no firm statement can be made regarding possible glacial advances within this period. Nor does data obtained so far deny the possibility of short lived pluvial events in the San Joaquin Valley (Croft, 1968), in the Lahontan system (Broeker and Kaufman, 1965) or in the Searles Lake system (Smith 1968). If such cooling or increased precipitation occurred, the soil evidence from the Sierra indicates that it should have substantially ended by 8700 years B.P.

The middle Holocene climate

Climatic indicators pertinant to the Pacific border suggest several intervals of warm, dry conditions within the period from 8700 to 2700 years B.P. Antevs (1948) originally recognized a warm, dry interval which he called the Altithermal. The temporal limits, geographical extent, and exact climatic parameters of the interval have never been determined.

In the present study the abrupt and synchronous change to gravelly-loam deposits and the presence of large mature conifers are taken to indicate a period of drier soil conditions best expressed in the middle unit at East Meadow. Minor woody peat beds, forest soils, and deposits of varied organic content within middle depositional units may reflect either episodes of cut and fill or minor paleoclimatic variations. A large number of radiocarbon dates would be required to determine whether there are hiatuses produced by cut and fill or whether some of the similar appearing layers in different meadows are representations of climatic events. The generally dry-soil conditions implied by the middle depositional unit is broadly interpreted to represent deposition under a climate characterized by reduced winter snowpacks unable to sustain high ground-water levels in valley bottoms.

From palynological study Adam defines a middle Holocene pollen zone containing radiocarbon ages 6990 and 2830 years B.P. and bounded by ages extrapolated on the basis of uniform sedimentation rates, of 8900 and 2400 years B.P. Variations among forest-genera pollen are minimal in this interval, but large variations in pollen from aquatic species are thought to reflect changing water levels in swamps (Adam, 1965).

Detailed lichenometric dating of Holocene glacial moraines in the Sierra by Curry (1968) led to recognition of several neoglacial advances since 2700 years B.P. and no advances in the interval 2700 to 6000 years B.P. A suspected glacial deposit overlaying a crushed root radiocarbon dated 7030 years B.P. (Curry, 1968 and 1970) has since been discounted as a landslide without paleoclimatic significance (Curry, 1971). Although no neoglacial advances have been discovered in the interval 8700 to 2700 years B.P., they may have occurred but were less extensive than the more recent neoglacial advances identified by Curry (1968). This point is important, because the Triple Lakes advance in the Colorado Front Range is a well-documented and most extensive neoglacial event that must lie between radiocarbon ages 3000 and 5000 years B.P. (Benedict, 1973). Carrara and Andrews (1973, p. 469-70 and 460-61) and Benedict (1973) show that significant Rocky Mountain advances are not necessarily synchronous with those in the Sierra: yet, the Sierra neoglacial events are recognizable in the Rockies.

A precise paleoclimatic record of the latter half of the Holocene is currently being obtained by LaMarche (1974) from changes of tree-ring width in bristlecone pines of the White mountains, 50 km east of the Sierra Nevada crest. The climate so inferred is one of relatively warmer and drier conditions from the beginning of record, 4000 B.C. (5200 B.P.)[5] until 110 B.C. (2900 B.P.)[5] except for one short period of cool moist conditions between about 3100 B.C. (4300 B.P.)[5] and 2800 B.C. (4100 B.P.)[5] (LaMarche et al, 1974). Calibration of tree-ring width anomalies with climatic parameters during the historical period of meteorological record will hopefully provide a quantitative measure of the past climatic parameters (LaMarche, 1974).

Additional paleoclimatic information is obtained by LaMarche (1973) from a fossil treeline in the White Mountains 150 meters above the present treeline. The oldest remnants represent trees established more than 7400 calendar years ago (about 6400 B.P.).[5] It is thought that the present upper treeline was established by cool climatic conditions of the last few hundred years. Using the present lapse rate, and the observation that timberline is established by the 10° C isotherm of the warmest summer month, LaMarche (1973) infers that the summer temperatures near timberline were on the average 1.5° C warmer during the Altithermal than during the last episode of climatic cooling since 1100 A.D. (900 years B.P.).[5]

Thus the chronology from Holocene glacial advances and studies of the Bristlecone pine indicate a warm period extending back to at least 6400 years B.P. The stratigraphic evidence from meadow deposits and from pollen would suggest extending this climatic interval back to about 8700 years B.P.

The Neoglacial Climate

The rebirth or renewed growth and all subsequent fluctuations of alpine glaciers after the time of maximum hypsithermal[6] glacier shrinkage following Wisconsin glaciation has been termed neoglaciation (Moss, 1951, and Denton and Porter, 1970). Neoglacial advances have occurred several times within the last 6000 years (Denton and Porter, 1970), yet it is not known with certainty whether most of these advances represent global phenomena, footnotes for page 121 and 122.

[5]Conversions of calendar years to radiocarbon years B.P. uses the curve of Suess (1970).

[6]The hypsithermal interval is defined by Deevey and Flint (1957) as the post-glacial interval when most of the world entered a period when mean annual temperatures exceeded those of the present.

or whether some are caused by local situations. Porter (1974) has reviewed the chronology of such advances from various mountain ranges in the temperate and polar regions of the World. There is broad agreement from widely separated ranges in both hemispheres for the significant neoglacial advances occurring within a few hundred years of 5300, 2800, 1200, and 300 calender years ago. The last event is a well-documented global phenomena. Evidence for the other advances is seemingly absent in some ranges, and it cannot be said with certainty whether this is due to 1) inadequate study and dating of moraines (Porter, 1974), 2) obliteration by a subsequent, more extensive advance, or 3) lack of an advance. For example, the 5300 year event is the most extensive in the Colorado Front Range (Benedict, 1973), yet it has not been recognized in the Sierra. Conversely, the major neoglacial advance c.a. 2800 years ago and the nearly as extensive advance c.a. 1200 years ago in the Sierra is only weakly expressed in the Colorado Front Range (Curry 1968, revised for Benedict, 1973).

Indicators of the approximate onset date of neoglacial climatic conditions in the Sierra are consistent. Lichenometric age of the most extensive advance was determined by Curry (1968) at 2700 years B.P. A pollen zone attributed to cooler climate in the Sierra is identified by Adam (1965), and the climatic change must be younger than a radiocarbon age of 2830 years B.P. Re-working the palynology of the same core, and extrapolating from that age, Zauderer (1973) concludes that the neoglacial-hypsithermal transition in the pollen record occurred at 2400 years B.P. in the Sierras.

The most detailed late-Holocene chronology of climatic variation pertinent to the Sierra is provided by tree-ring studies of the bristlecone pine (LaMarche, 1974). Careful selection of trees at the upper treeline, sensitive to summer temperatures, and of trees at the lower forest limit, sensitive to soil moisture, provides detailed information on relative changes in both temperature and precipitation. Climate in the White Mountains became sharply cooler and moister 2700 years B.P.[5] and endured until 2150 years B.P.[5], and likewise starting in 1600 B.P[5] and extending to 1000 years B.P.[5] A more continuous, but less severe, cooling spanned the period 750 B.P. to about 150 years B.P. (LaMarche, 1974).

Radiocarbon chronology for the onset of increased soil moisture inferred from stratigraphy at meadow sites is tabulated in table 3-4. Best estimate of the rise of ground water level at East Meadow is 2500 years B.P. and at Tuolumne Meadows, 2300 years B.P. These estimates agree well with the onset of neoglacial conditions dated by Curry

TABLE 3-4. Bracketing radiocarbon ages for the onset of increased soil moisture conditions at meadow sites.

site	maximum age	minimum age
East Meadow	2830 B.P.	1545 B.P.
Tuolumne Meadows	1200 B.P.	3930 B.P.
Cabin Meadow		1200 B.P.
West fork, Long Meadow Creek	1210 B.P.	720 B.P.
Boggy Meadow		1200 B.P.

(1968), Adam (1967), Zauderer (1973), and LaMarche (1974). It is difficult to account for the meadowing of these sites in any other way than by climatic change. Meadows that lie farther south, Cabin Meadow, Long Creek Meadow, and Boggy Meadow, do not show such striking coincidences, the first two appearing to have "meadowed" at 1200 years B.P. and Boggy Meadow between 1200 and 720 years B.P., closer to 720 years. These dates fall within cool, moist periods identified by LaMarche (1974). On this basis, it appears that climatic variations govern the height of the growing-season water table. A high water table is characterized by the greater winter snowpack of cool-moist episodes, the water levels being maintained by infiltration of sustained meltwater discharges from snowbanks that do not disappear until late summer.

Origin of Meadows

Previous discussions of mountain meadow origins (Muir, 1894, Bradley, 1921, Sumner and Leonard, 1948, and Bakker, 1972) attempt to fit them into an orderly succession of landforms and plant community developments. The usually advocated origin is that meadows occupy the sites of glacially formed lake basins, which as they progressively filled with sediment went through a hydroseric succession of lake, marsh, meadow, and ultimately will reach the climax forest stage. As Muir recognized, the infilling of glacial lakes is a slow process, and it is only the shallowest of these lakes, filled principally by sand deltas from inflowing streams, that can have become meadows since the last glaciation. Muir writes that once it is established, " -- many a fine meadow favorably situated exists in almost prime beauty for thousands of years." The stratigraphic investigations of the present study show that few meadows are underlain by lake deposits and that the majority of meadows do not fit into any simple scheme of hydroseric succession. Many meadows lie in montane valleys that were never glaciated, and the surface water has always flowed through these sites and has not been impounded in closed basins.

The matter of meadow origin requires identification of processes which maintain an open meadow under the seemingly constant pressure of invasion from the surrounding forest. Some investigators conclude the origin of meadows from observations of present vegetation and surficial topsoils, such as in the northern Rocky Mountians, Griggs (1938) felt that fire maintains the openings, and Kuramoto and Bliss (1970) believed that fire both creates and maintains the subalpine meadows of the Olympic Mountains, Washington. In the northern Cascades, Brink (1959) and Douglas (1972) attribute maintenance of meadows to deep and long lingering snow cover in the meadow area relative to earlier melting snow in the forests.

Franklin et al (1971) studied subalpine meadows in Paradise Valley on Mount Ranier, Washington. These meadows here established upon deposits of the 6000-year old Paradise lahar identified by Crandall (1971). Franklin et al (1971) indicate that the meadows have never been forested in post-glacial time since neither charcoal nor rotten wood is found in soils except near tree clumps. They conclude that fire has had little or nothing to do with creation and maintenance of these meadows. Investigations in the Bighorn Mountains of Wyoming (Despain, 1971) and in the Jackson Hole area (Loope and Gruell, 1973) reveal little incipient conifer invasion of meadow areas, and these authors conclude that it is edaphic conditions that maintain the meadows.

Stratigraphic evidence of former forest intervals beneath existing bogs and peatlands are reported in the northeastern United States by Deevey (1958), in north-central U.S. by Heinselman (1970), in northwestern Russia by Neustadt (1967), and in Europe as discussed in Moore and Bellamy (1974). The modern swampy condition can perhaps be explained in some instances by a water-table rise associated with change toward a wetter climate. Sharsmith (1959 and 1962) suggested some meadows in the Sierra develop upon stream deposits as the water table rises. This suggestion seems the only plausible means by which meadows are established upon forested tracts and stream deposits in the Sierra.

Role of climatic change

It is proposed that many meadows do not evolve in a slow successional manner, but are created abruptly by changing hydrologic conditions associated with climatic variation. In the Sierra, topography and surface-water hydrology determines whether or not fine-grained soils are deposited and retained in stream valleys. Given a certain hydrologic regime, it is the stream gradient, the upstream drainage area, and to some extent the width of the valley floor that determine the competency of flow. In order to build a graded alluvial flat capable of harboring a high water table, the relationship between stream gradient and the upstream drainage area must lie within certain limits which are defined broadly in the Sierra by the domain on a slope-drainage basin plot (figure 2-1) where meadows occur. Outside these limits erosion does not permit aggradation in the valley, and no graded surface forms. Once the graded surface is formed it is the shallow groundwater system that determines whether or not a meadow can develop. Groundwater levels under such flats are maintained primarily by seepage from the valley sides and the stream inflow. The groundwater stored on the valley sides is annually recharged during the period of snow pack. If, prior to the current period of relatively cool and wet climate, the amount of snowfall was less, then the snowpack on the slopes at the end of winter would also be less and they would become snow free a month or more earlier. Short seasons of ground water recharge over several decades should diminish storage on the valley sides, and prior to the neoglacial, snowmelt was unable to maintain a high water table in the valley fill. Computations, presented in Appendix B, suggest that the summer flow of ground water from the valley sides to the upper perimeter of the flat at East Meadow must have nearly doubled in order to raise the water table two feet and change this formerly forested site to a wet meadow.

Great variability currently occurs in the winter snowpack in the Sierra, ranging from 300 per cent to 40 per cent of the average (California Department of Water Resources, 1960-1974). In years of reduced snowfall the montane basins can be snowfree by the end of April but in years of heavy snowfall, the snow lingers well into July. Curry (1968) concludes from an analysis of the historical snowfall record and the downslope extent of neoglacial moraines in the Central Sierra, that 60 consecutive years of 120 to 130 per cent snowfall can account for all neoglacial advances in the last 2600 years. It is reasonable to expect that the period prior to the neoglacial was characterized by many consecutive years of reduced snowfall. Reduction in snowpack required to reduce the pre-neoglacial summer ground-water flow at East Meadow by fifty per cent, can only be guessed at. Perhaps sixty to eighty per cent of the present snowfall for several decades would be adequate. Monitoring of late summer water table levels for many consecutive years in typical meadow is required to evaluate the long-term relationship of ground-water to snowfall in these basins. The meager two year record from the Central Sierra Snow Laboratory (figure 2-7) is not sufficient. These data indicate only that the water stored annually upon the valley sides is enough to restore the water table to the surface within two months after the summer evapotranspiration deficit has lowered the water table in the meadow by 2 feet.

A sequential history of mountain meadow development, controlled by climate, is outlined in figure 3-31. This scheme broadly applies to most upper montane meadows, although certain features may be absent. Not all alluvial flats will have exactly the same sensitivity to climatic fluctuation. Many smaller meadows may only show the late Holocene events.

Lodgepole pine invasion of meadows

Loss of meadow acreage to forest invasion has been a troublesome problem to those charged with managing the mountainous areas of the western United States. A remarkable feature of the conifer invasions is that the establishment of invading conifers is confined to short intervals of a few years, resulting in remarkably even-aged stands. These establishment intervals are often duplicated in widely separated meadows throughout the range. By far the most common encroaching tree in the Sierra is Lodgepole Pine. Most montane and subalpine meadows are surrounded by a relatively dense growth of lodgepoles, some of them a few hundred years in age. It is common to find a clump of mature Lodgepole Pine well out in the meadow associated with a large rock or a better drained spot. Also common are small, stunted, lone, lodgepoles in wetter parts of the meadow. Most curious, however, are the even-aged stands, often a few acres in extent, that have recently extended from the perimeter into the meadow.

The Lodgepole Pine is more tolerant of wet soil than other Sierra conifers, but it is not resolved whether it is the inability of tree roots to compete with meadow-community plants for nutrients and space, or if it is the anaerobic soil conditions that are unfavorable.

Few invading stands of lodgepole have been dated. Leonard et al (1968) report an invasion in the Rock Creek area of Sequoia Park between 1906 and 1911 based on increment borings. Bennett (1965) reports common establishment years of 1924, 1931, and 1939 for lodgepoles invading Vidette Meadows of Sequoia Park. Sharsmith (1962) notes that massive invasions occurred in Tuolumne Meadows of Yosemite Park just prior to 1920 working from old photographs, and in another meadow between 1903 and 1905.

Lodgepole pines quickly invade abandoned roads, trails, and ground formerly disturbed by livestock. This observation led Sharsmith (1972) to suggest that overgrazing in the late 19th and early 20th century may have been responsible for historical invasions prior

1. During the late Wisconsin ice age, peak flows of the stream are sufficiently competent to transport coarse alluvium along a 1 to 2 per cent stream gradient and out of the valley. Deposits are limited to bed-load cobbly gravels. The watershed is sparsely vegetated. Timberline is near the present upper montane belt.

2. Early post-glacial warming, just prior to 10,200 years B.P. , is accompanied by montane-belt forestation of the watershed. Diminished peak flows allow deposition of fine clastics and organic material in the stream channel. Willow and sedge meadows and bogs form in low areas. Soil profile develops on riparian forest sites.

3. Intervals of diminished winter snowpack begin about 8700 years B.P. Low summer water tables allow forest invasion of bogs and meadows. Continued low snowpack during the middle Holocene maintains the forest environment on the valley bottom. Forest soil and slope wash alluvium accumulate on the valley floor.

4. Extremes of neoglacial climate about 2500 and 1200 years B.P. cause late melting snows and a rising water table leading to mortality of the forest on the flat and establishment of open meadow conditions.

5. Late melting snows of the neoglacial sustain high ground water and meadow conditions. Organic topsoils, peats, and sheets of flood gravel aggrade the surface. Aggradation on a protective cover of meadow sod leads to oversteepening of the surface slope.

6. Summer use of meadow by livestock damages the protective sod. During winter plant dormancy, flood waters from torrential winter rain storms initiate gullies leading to erosion of the valley fill and destruction of the meadow.

Figure 3-31. The developmental history of a montane meadow interpreted from the stratigraphy of deposits in valleys on the west slope of the southern Sierra Nevada, California.

to 1920.

It is noteworthy that in the stratigraphy beneath meadows, spanning much of the Holocene, only in the Cabin Meadow section is there a stump level of exclusively Lodgepole Pine. In other sections, other pine and fir species of a mixed conifer forest are represented. All the stump lines are imbedded in a forest soil, and not wet meadow soils. Also not more than one or two such stump lines in a section suggesting that invasion of meadows has not been frequent in the past.

A few ailing or dead stands have been seen on the edges of modern meadows; one in particular at Snow Corral Meadows near Dinkey Creek. In the Tuolumne Meadows area some have been clearly associated with infestation by the needleminer moth (Koerber, 1973). Others may be caused by senescence or small water-table changes. Some process must naturally keep the numerous seedling trees in check, and it is likely that fluctuating water tables prevented invasion in the past and continue to do so.

Franklin and others (1971) studied invasion by fir and hemlock of subalpine meadows in the Cascade Range. A period of vigorous invasion is documented between 1928 and 1937. They correlate this interval with a period of negative mass budget of the Nisqually Glacier on Mt. Ranier (Meier, 1965) related to a much diminished and early melting snowpack. They suggest that infrequent coincidence of a good seed crop in the fall with early melting snow the following spring characterizes intervals of tree establishment. This explanation may also apply to the Sierra, but a comparative study is yet to be done.

A method of finding the age of seedling Lodgepole Pines is to count the top shoot and successive annual branch lines. In the summer of 1974, the branch ages of 57 seedlings recently established in the meadow at Tuolomne Ranger Station were counted. They ranged from the summer of 1964 to the summer of 1972, but by far the commonest years, in order of abundance of seedlings were 1968 (19 seedlings), 1970 (13 seedlings) and 1967 (10 seedlings). Records show that 1968 and 1970 were abnormally low snowpack years (less than 80 per cent), and most montane sites were snowfree in early May. In 1967, however, heavy snowpack and snow lingered until mid-June. There is thus a suggestion that most seedlings are established in years of low snowpack.

CHAPTER IV

MONO AND INYO CRATER ERUPTIONS – DISTRIBUTION, CORRELATION AND RADIOCARBON DATING OF LATE HOLOCENE TEPHRA

Introduction

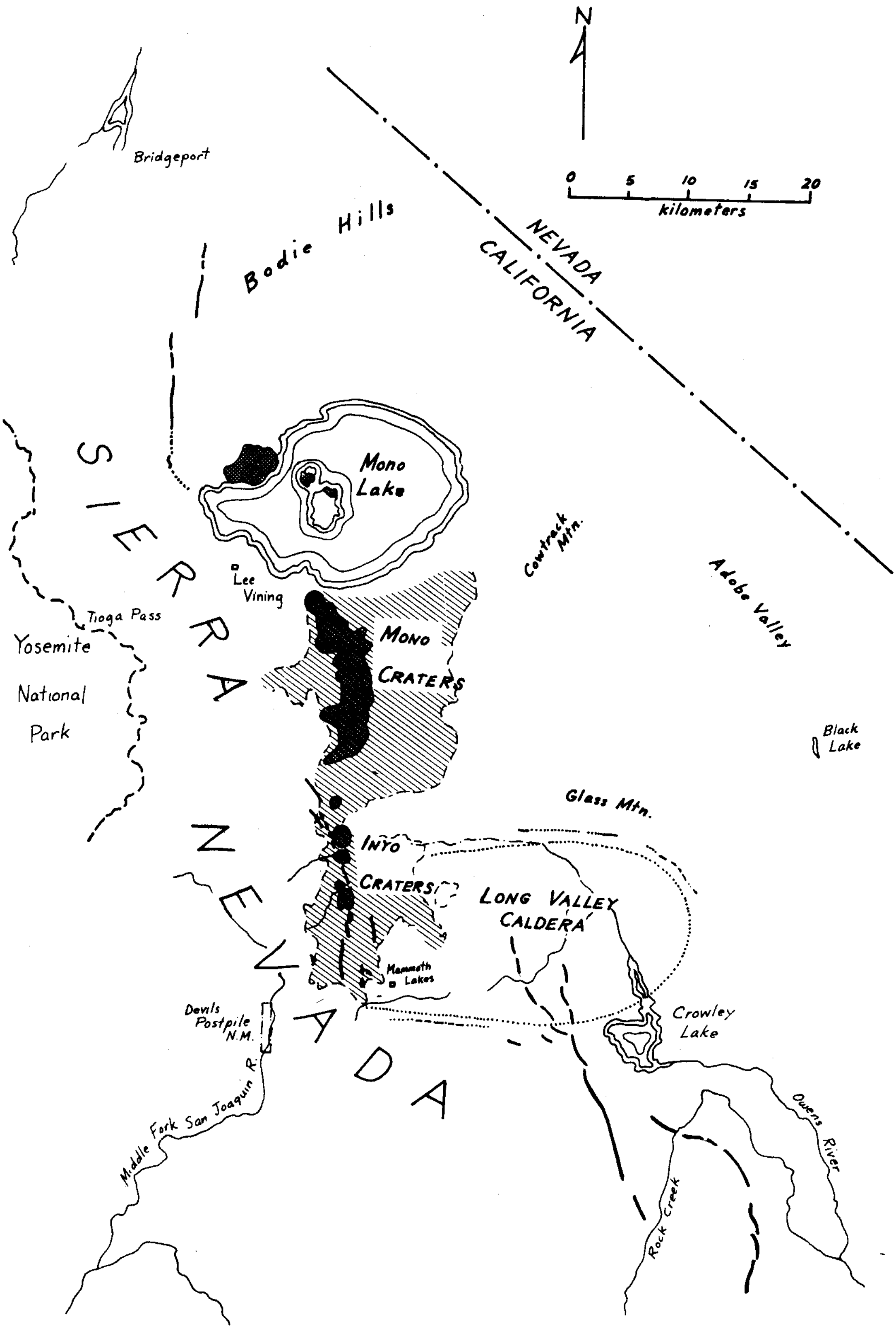
The Mono-Inyo Crater chain is a sequence of late Quaternary domes, flows, craters, and tephra deposits, dominantly rhyolitic, that erupted along a 40-kilometer arc at the eastern base of the Sierra Nevada. Holocene volcanic features extend north from the 680,000-year old Long Valley caldera (figure 1) currently being investigated by Bailey (1973a) and Bailey and others (1973). Studies have been made of individual groups of features within the Mono-Inyo chain (Smith, 1973; Loney, 1968; Kistler, 1966; Putnam, 1938; Balk, 1937; Mayo and others, 1936; and Williams, 1932), but the entire complex has not been systematically mapped, so temporal relationships are not fully understood. This paper reports new radiocarbon based ages for eruption of the two most recent tephra layers, their distribution in the Sierra Nevada, and their correlation to tephra-ringed eruptive centers in the chain, as indicated by trace-element content.

The eruptive sequence

A sequential development of the eruptive forms of the Mono-Inyo Crater type has been outlined by Smith (1973). The initial stage is an explosion crater rimmed by tephra. This is followed by extrusion of a dome on the crater floor, and if extrusion continues, it overflows the tephra rim to form a coulee flow that extends several kilometers from the vent. In some instances the dome itself may be subsequently cratered by explosion or collapse and invaded by yet another dome which in turn may also be cratered by collapse or explosion producing a double crater. The eruptive sequence sometimes commences with the extrusion of pumiceous obsidian domes or flows without the preceding violent pyroclastic eruptions. It is the initial explosive stage, however, that supplies large volumes of pumiceous tephra.

Direct observations of an eruptive sequence producing forms seen in the Mono-Inyo

Figure 1. Map showing Holocene volcanic rocks (cross hatched pattern), Holocene pumice cover greater than 15 feet thick (diagonal line pattern), the bounding faults of the Pleistocene Long Valley Caldera, and faults along the Sierra escarpment with probable Holocene displacement. (From the Geologic Map of California, Walker Lake sheet, 1963 and Mariposa sheet, 1967, and K. Lajoie, 1968.)


N
0
5
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kilometers
NEVADA
CALIFORNIA
Bridgeport
Bodie Hills
SIERRA NEVADA
Mono Lake
Lee Vining
Tioga Pass
Yosemite National Park
Cowtrack Mtn.
Adobe Valley
MONO CRATERS
Black Lake
Glass Mtn.
INYO CRATERS
LONG VALLEY CALDERA
Mammoth Lakes
Devils Postpile N.M.
Crowley Lake
Middle Fork San Joaquin R.
Owens River
Rock Creek

chain have not been made, however, events occurring during the 1912 eruptions in the Valley of Ten Thousand Smokes, Alaska (Fenner, 1923 and 1950, and Curtis, 1968) are enlightening because the major volcanic vent, Novaputra, resembles in size and shape the tephra-ringed dome, Panum Crater, on the south shore of Mono Lake (Green and Short, 1971). The Novaputra vent erupted a volumnous (10.8 km^3) incandescant tuff-flow followed by several equally volumnous eruptions of pumiceous tephra totalling 19.8 km^3 of expanded tephra. The resulting tephra-ringed crater was then invaded by a viscous plug of banded pumiceous obsidian 400 meters in diameter and 100 meters high. Major eruptive activity commenced on June 6, and had entirely ceased by the end of July, 1912. The area has been dormant since that time. If similar appearing features in the Mono-Inyo chain form in a manner similar to Novaputra, ages on tephra should agree closely with the age of domes or flows of the same eruptive sequence.

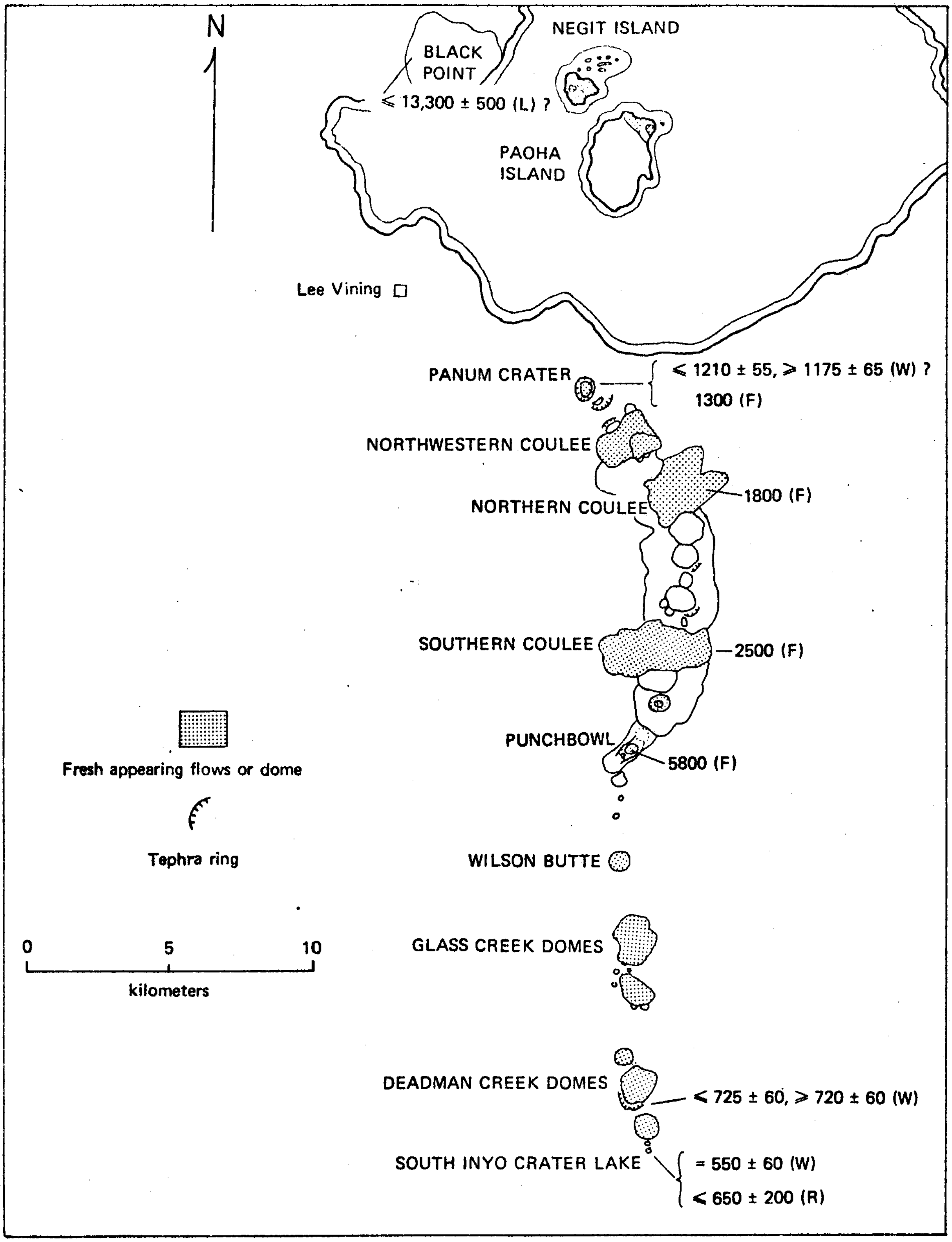
Previous work

Geochronology of the volcanoes: Dalrymple (1967) used samples from the chain to test the applicability of K-Ar age-dating to Holocene volcanics. The analysis had poor reproduceability owing to insufficient radiogenic argon, but they indicated ages of the order 10,000 years or less.

Friedman (1968) dated obsidian samples from the northern Mono Crater group by the hyrdration-rind method. Lacking radiocarbon calibration of the hydration diffusion rate for the Mono Crater area, he used a curve from Medicine Lake, northern California, which yielded late Holocene ages on several Mono Crater obsidian bodies (figure 2). The method gives promise of establishing relative temporal relationships within these volcanoes, and if the hydration rate can be locally calibrated to a few radiocarbon or dendrochronologically dated materials, absolute ages may be possible.

A radiocarbon based age between 500 and 850 years is reported by Rinehart and Huber (1965) for the steam explosion that excavated the south Inyo Crater Lake. They also report a maximum radiocarbon age of 1600 years for the eruption of pumice lapilli that blankets the Mammoth Lakes area, but they do not identify the vent from which the pumice erupted.

Batchelder (1970) reports that 5 pumiceous tephra layers were found in piston cores of the organic lacustrine silts of Black Lake, in Adobe Valley, 10 miles east of the volcanic



Compilation of radiocarbon and hydration rind ages on the Mono and Inyo Craters. Map based on Lajoie (1968), Smith (1973) and reconnaissance mapping. Source of radiocarbon ages: R = Rinehart and Huber (1965), L = Lajoie (1968) and Denham and Cox (1971), W = this study; Hydration-rind ages (F) = Friedman (1968).

Figure 2

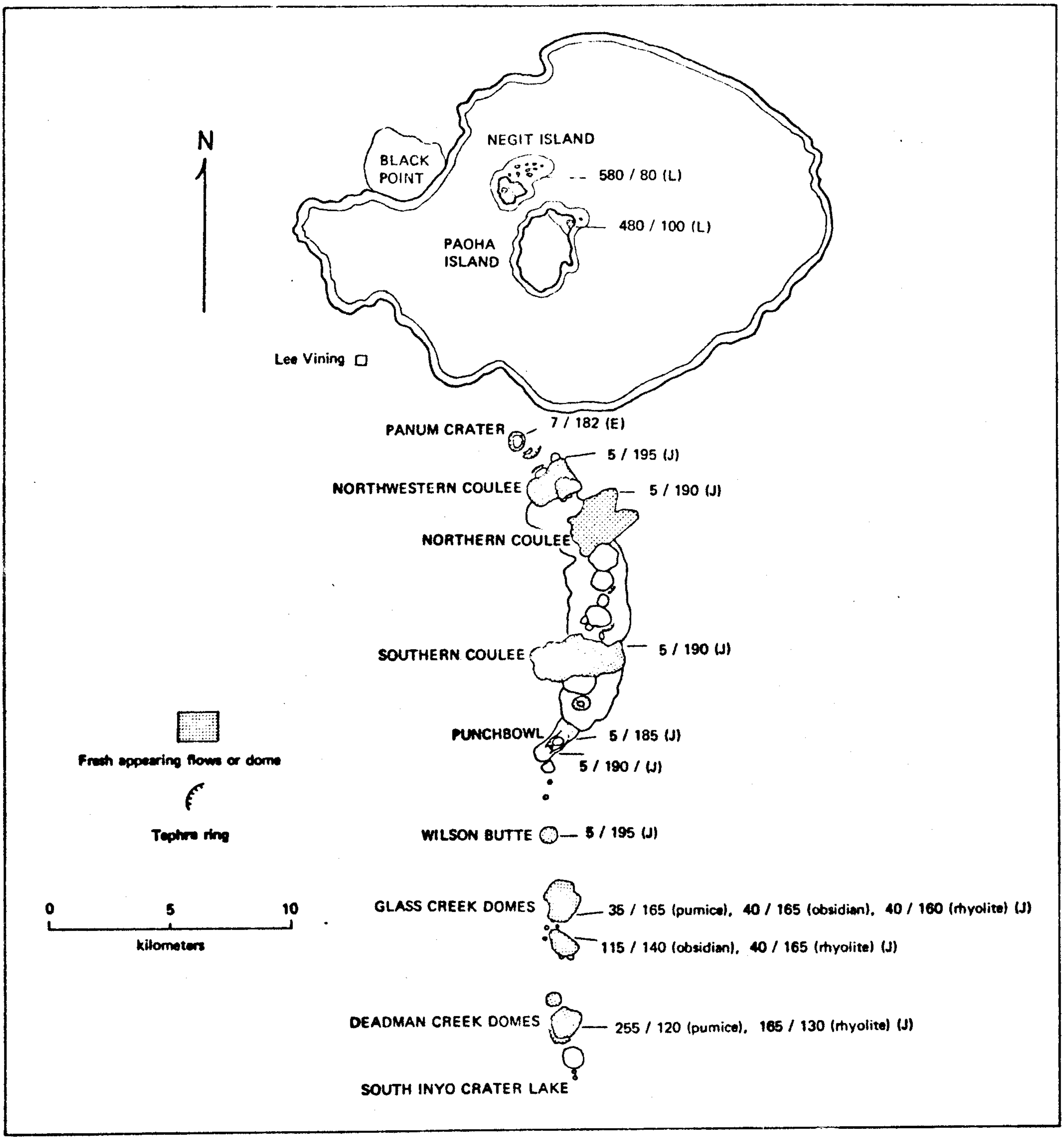


Figure 3. Compilation of Strontium and rubidium analysis on extrusive rocks of the Mono and Inyo Craters: Sr/Rb values in ppm followed by data source: J = Jack and Carmichael (1968), L = Lajoie (1968), (X-ray fluorescence analysis); and E = T.O. Early, unpublished (isotope dilution).

chain. Two layers are younger than radiocarbon age 2190 years B.P. and three layers occur between sediment dated 4588 and 5230 years B.P. No tephra is reported from the sediment between 5230 and 11,350 years B.P.

Lajoie, (1968) studied 17 rhyolite tephra layers in the lacustrine silts deposited during the late Wisconsin high stand of Mono Lake. Ages of the 17 tephra layers based upon extrapolations from 2 radiocarbon dates (Table 1) range from 13,100 to 30,310 years B.P. (Denham and Cox, 1971). Two basaltic tephra layers occur in the section. The most prominent basaltic tephra is dated 13,100 and correlated by Lajoie (1968) to the Black Point basaltic cone which erupted sub-aqueously (Christensen and Gilbert, 1964) during a high stand of Mono Lake. He was unable to identify the vents from which the rhyolite tephras erupted, but trace-element content indicates a geochemical affinity with the northern Mono Craters.

Trace-element geochemistry of the volcanics: The trace-element chemistry (Jack and Carmichael, 1968, and Lajoie, 1968) of the northern group of Mono Crater bodies, excepting those on Paoha and Negit Islands in Mono Lake, are uniform and homogeneous. Trace-element chemistries of the southern Inyo group of eruptives and recent extrusives on the islands are distinctly different and seemingly unique for each eruptive vent, particularly in Zr, Sr, and Ba content.

In a compilation of existing data for Rb and Sr analyses on the extrusive materials (figure 3), Sr shows large variation throughout the volcanic chain with values ranging from 5 to 580 ppm: the Sr content of the Inyo Craters and the island volcanics appears to characterize each eruptive center. The Mono Craters complex, including Wilson Butte, have remarkably low Sr content (less than 10 ppm) and indistinguishable trace-element chemistries. Rb content variations through the volcanic chain are not large: all of the analyses lie between 80 and 195 ppm. Analyses of pumice, rhyolite, and obsidian from the same vent suggest that trace-element chemistry does not vary greatly throughout an eruptive sequence, although many more analyses of this type are needed to ascertain the reliability of this conclusion.

Seventeen rhyolitic tephra layers in a section of late-Wisconsin Mono Lake sediments were analyzed by Lajoie (1968) and found to have a trace-element chemistry matching the Sr-poor northern Mono Crater bodies. No tephra was found in that section with the higher Sr content characteristic of the Inyo Craters and the Mono Lake island volcanics.

Table 1a. Radiocarbon ages associated with tephra layers

Locality	Material	Age	Radiocarbon Designation	Reference
Wilson Creek Formation	ostracod tests	13,300 ± 500 23,300 ± 300		Lajoie (1968) and Denham and Cox (1970)
Black Lake Adobe Valley	organic sediment	5230 ± 110 4580 ± 130 2190 ± 90	GAK-1850 GAK-1852 GAK-2476	Batchelder (1970)
Mammoth Mountain	charred wood	1440 ± 200	W-727	Rinehart and Huber (1965)
Mammoth Mountain	charcoal and wood fragments	920 ± 80	UCLA-908	Berger and Libby (1966)
Bodie Creek	carbonized wood	700 ± 200	W-629	Rubin and Alexander (1960)
Inyo Crater Lakes	conifer log	650 ± 200	W-1431	Rinehart and Huber (1965

Table 1b. New radiocarbon ages associated with tephra layers

Locality (refers to figure 6)	Material	Age	Radiocarbon Designation	Reference
A: South Inyo Crater Lake	pine log	710 ± 60	UGa-603	Brandau and Nokes (1974)
B: Lower Cabin Meadow, near Dinkey Creek	fir log	760 ± 60	UGa-602	"
C: W.Fk. Long Meadow Ck.,	fir stump	1210 ± 55	UGa-405	"
Wishon Reservoir	fir log	1175 ± 65	UGa-451	"
D: Devils Post Pile Rd. Middle Fk. San Joaquin	charcoal	3375 ± 140	UGa-449	"
E: East Meadow Yosemite N.P.	charcoal	1545 ± 90	I-6049	Wood (1972)

Eruption of rhyolitic tephra with high Sr content appears to be restricted to activity within the Holocene since such tephra are absent in late-Wisconsin sediments.

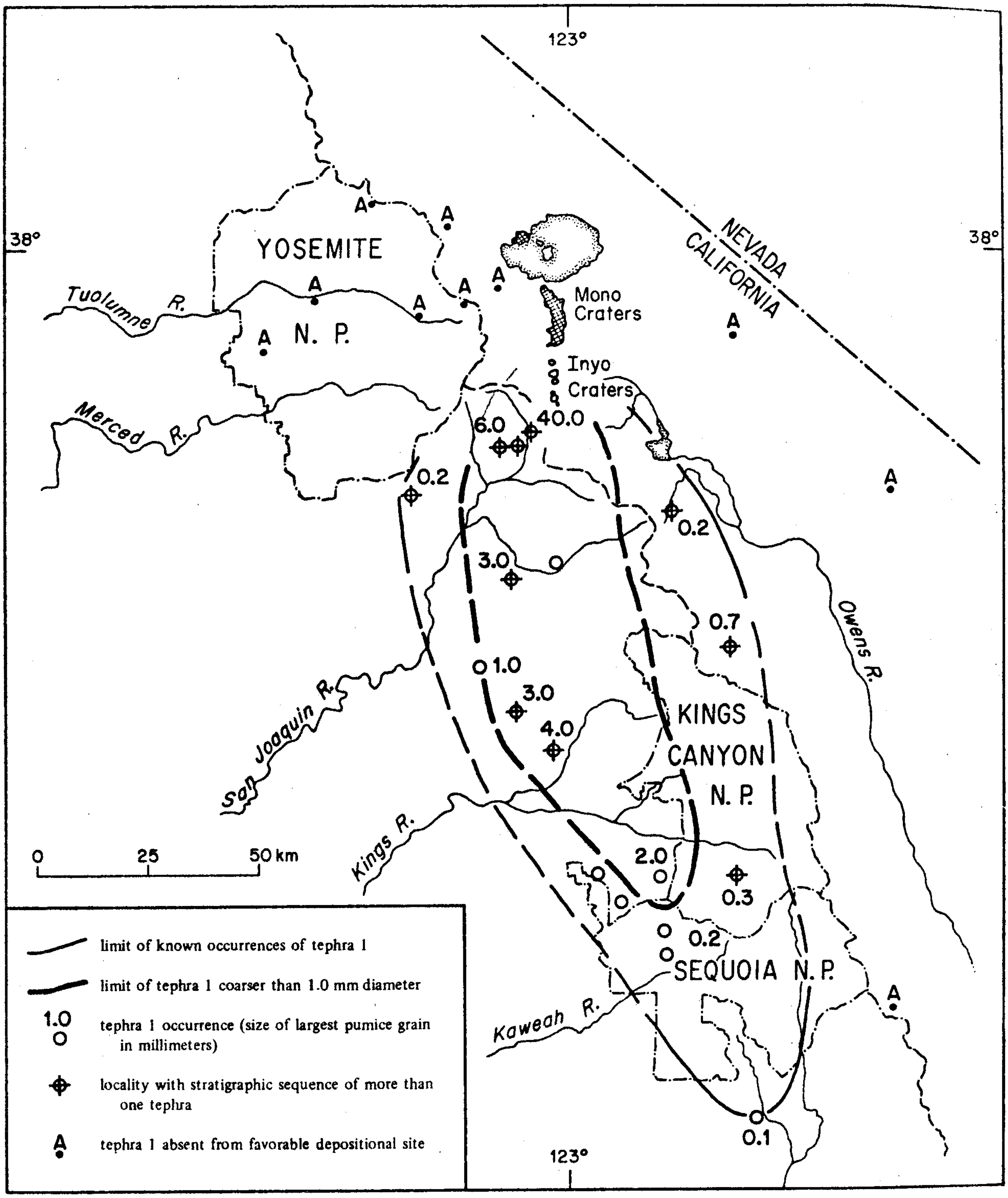
Distribution and Character of Tephra Layers 1 and 2

Two recent ash and lapilli layers, a few centimeters thick, are recognized in meadow topsoils of the southern Sierra Nevada at widely distributed localities (figures 4 and 5). These are the only significant tephra layers found within meadow-deposit sections on the west Sierra slope between latitudes 36^{o} and 38^{o}, excepting two deeper and older tephra layers (figure 8) recognized only in the upper San Joaquin River drainage.

Distribution of tephra layers

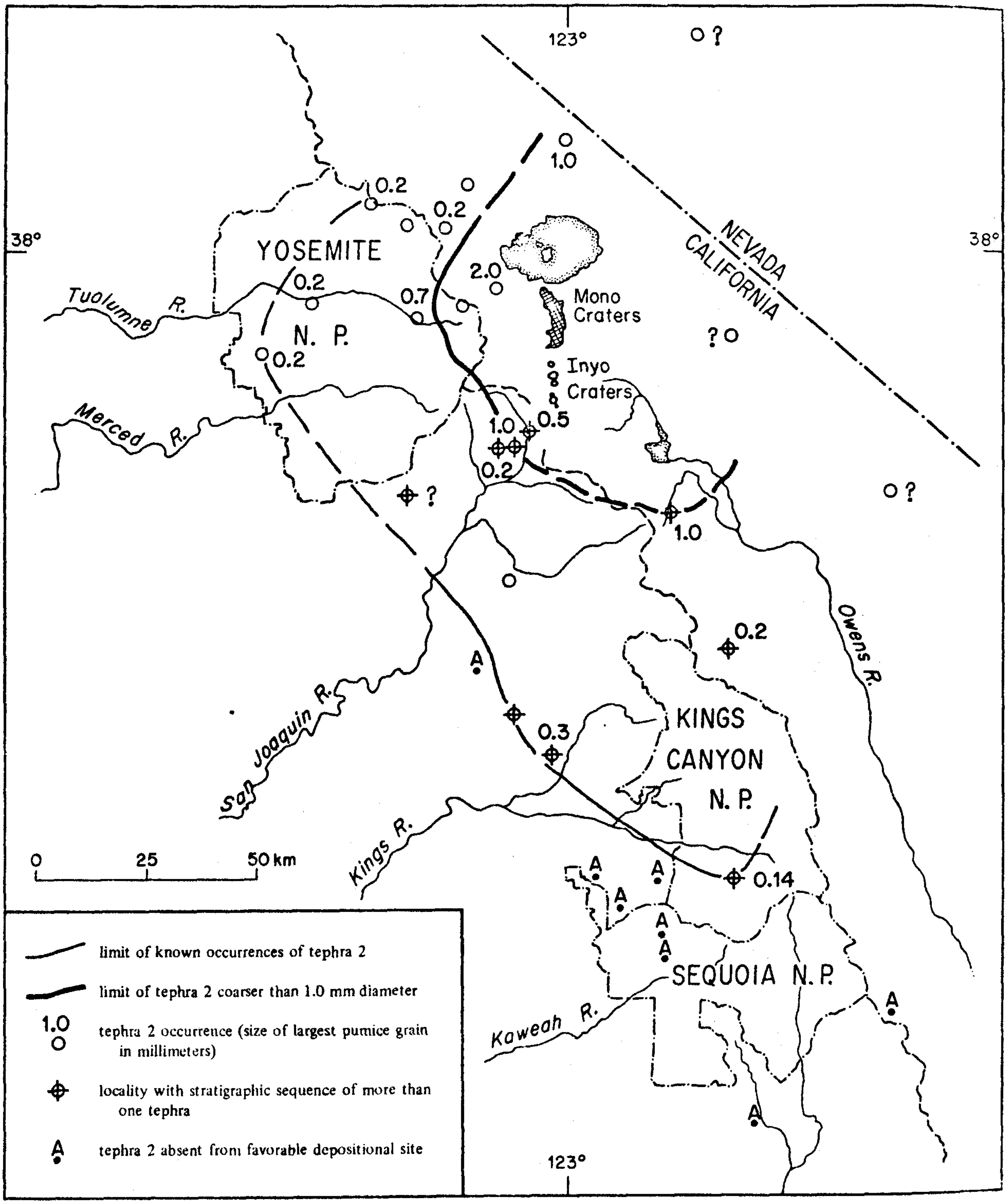
Tephra 1: The distribution of remnants of the youngest tephra defines a south-blown lobe extending 120 miles (190 km) from the Inyo Craters (figure 4). This is the tephra that forms the thick pumice mantle in the Mammoth Lakes area, and it can be traced from the upper San Joaquin drainage south to the valley of Little Kern River. It is probably the volcanic ash that F.E. Matthes found in bedrock potholes in Sequoia National Park (Stewart, 1929, and Elsasser, 1965). Pumice Lapilli that Jahnda (1966) reports upon all glacial moraines in the Ritter Range, excepting those of the Matthes age neoglacial advances, to which Curry (1971) assigns lichenometric ages from 55 to 620 years B.P., is probably also tephra 1. Pumice fragments, concentrated in natural depressions and increasing in abundance northward as reported in the eastern half of the Mt. Abbot and in the western half of the Kaiser Peak quadrangles (Lockwood and others, 1972) also probably represent this tephra. At more distant sites it commonly occurs as a thin layer, 0.5 to 1 foot deep in mountain-meadow deposits.

Tephra 2: Distribution of tephra layer 2 in the Sierra (figure 5) does not delineate a lobe. The principal lobe of transport may lie eastward in the less well investigated Great Basin Area. Several tephra occurrences are known to the east, but they have not yet been positively correlated with tephra 2. In the Sierra, tephra 2 commonly occurs 1 to 2 feet deep in meadow deposits from northernmost Yosemite to the south side of Kings Canyon. Near the north end of the Mono Craters it forms a one-foot bed of white ash in colluvial deposits in Lee Vining Canyon and in the Bodie Hills.



DISTRIBUTION OF TEPHRA LAYER 1, AGE = 720 YEARS B.P.

Figure 4.



DISTRIBUTION OF TEPHRA LAYER 2. AGE = 1200 YEARS B.P.

Figure 5.

Textural differences between the tephra layers: Textural variation within the tephra units is indicated (figures 4 and 5) by a contour delineating the area with fragments coarser than 1 millimeter. Size of the largest pumice fragment in the layer at each locality is annotated to indicate the principal directions of transport and to demonstrate the significant textural difference between these two tephras. This may reflect a difference in the nature of the eruptions which produced a predominantly coarse lapilli for tephra 1 and a fine ash for tephra 2. Consideration of the textural variations (figures 4 and 5) and a limited amount of isopach data suggest that tephra 1 erupted from the southern part of the volcanic chain and tephra 2 from a source near the northern end.

Mineralogy and petrology of the tephra: Differences in the microphenocryst content of these tephras has been reported (Wood, 1972). Tephra 1 has square prismatic, often twinned, microphenocrysts of sanidine that comprise about 30 per cent of the grains examined in a mount under crossed-nichols. Tephra 1 also gives a strong 3.27 A^{o} X-ray diffraction peak of sanidine. Other grains and the matrix associated with the microphenocrysts are pumiceous glass. About 30 per cent of the pumice grains have vesicles sheared into elongate pipes. In the remainder, the vescicles are round or oval.

Tephra 2 is devoid of microphenocrysts although powder X-ray diffraction shows a weak 3.37 A^{o} peak of biotite. Over 50 per cent of the pumice grains in tephra 2 have vesicles sheared into elongate pipes. Refractive indices of the glass in both tephras ranges from 1.496 to 1.500, and it is not a useful criteria for distinction between these tephras.

Preliminary microscopic examination of the two older tephra layers in the upper San Joaquin drainage (figure 8) revealed no microphenocrysts; therefore, the square prismatic sandine microphenocrysts appear to be diagnostic of tephra 1 in the southern Sierra.

Trace-element analyses: Rhyolitic glass from the tephras 1 and 2 have been analyzed by rapid scan X-ray fluorescence for trace amounts of Sr, Rb, and Zr. Analytical methods are discussed in appendix. Results are shown in table 2.

The large difference in Sr content and the considerable difference in Zr content would seem to provide a firm basis for distinguishing between these two tephra layers based solely on their trace-element content.

These trace-element data, in addition to the stratigraphic, mineralogic, and textural considerations, establish a means of correlation for discontinuous occurrances of these

Table 2. X-ray fluorescence analyses in ppm

Locality (no. refers to figure 6)	Sr	Rb	Zr	Tephra layer
(4) East Meadow, Yosemite N.P.	5 ± 3	195 ± 10	125 ± 30	2
(3) W. Fk. Long Meadow Ck. Sierra N.F.	20 ± 10	206 ± 10	120 ± 10	2
(1) Starkweather Lake flat, Devils Post Pile N.M.	225 ± 15	205 ± 10	400 ± *	1
(2) Boggy Meadow, Kings Canyon N.P.	210 ± 15	204 ± 10	375 ± *	1

Uncertainty refers to estimated precision of repeated analyses on samples and standards.

* linear extrapolation of calibration curve for high Zr.

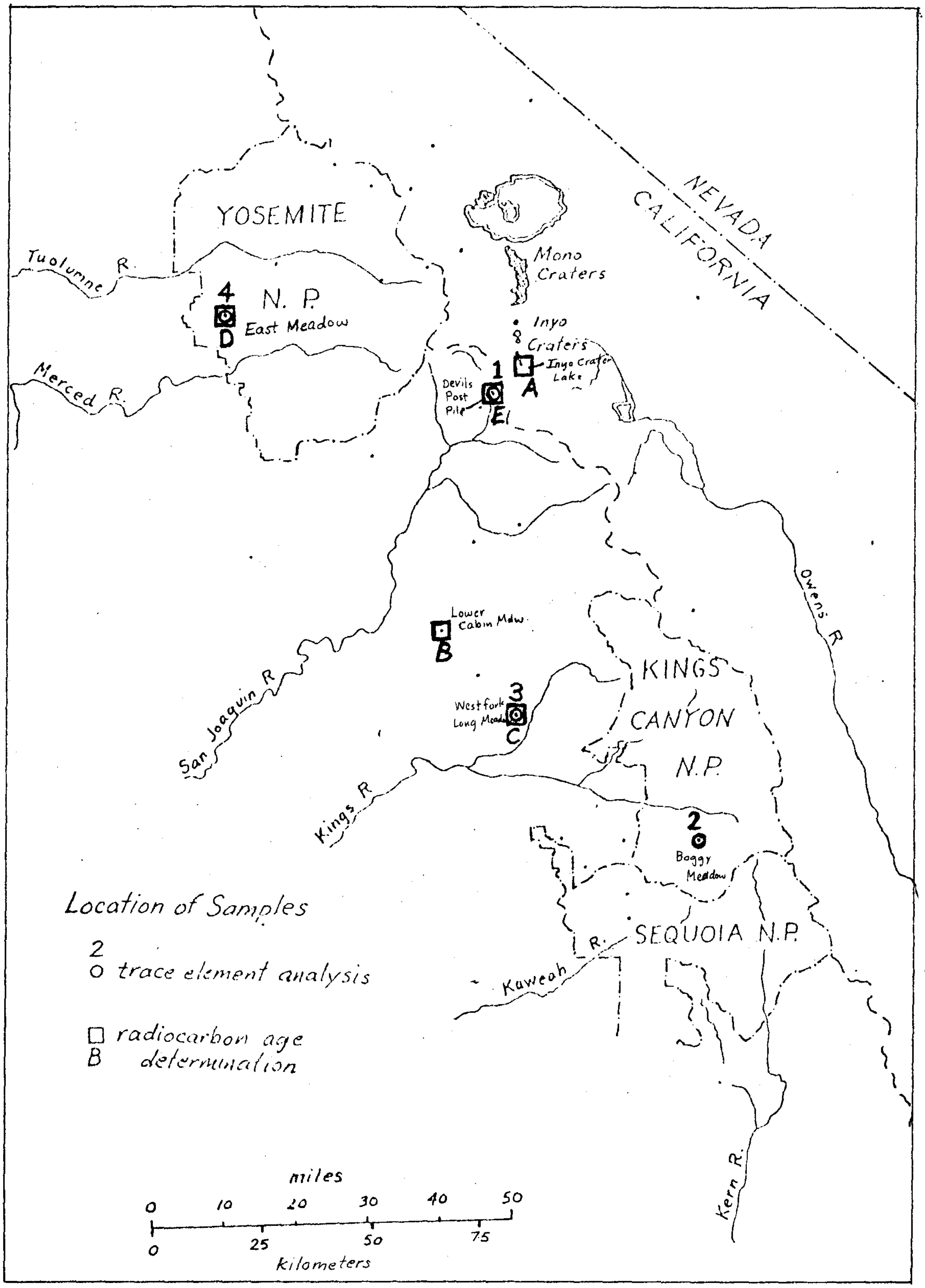


Figure 6. Location map for radiocarbon dated samples (Table 1b) and trace element analysis on tephra (Table 2)

tephras in eastern California.

Correlation of Tephra to Their Eruptive Vents

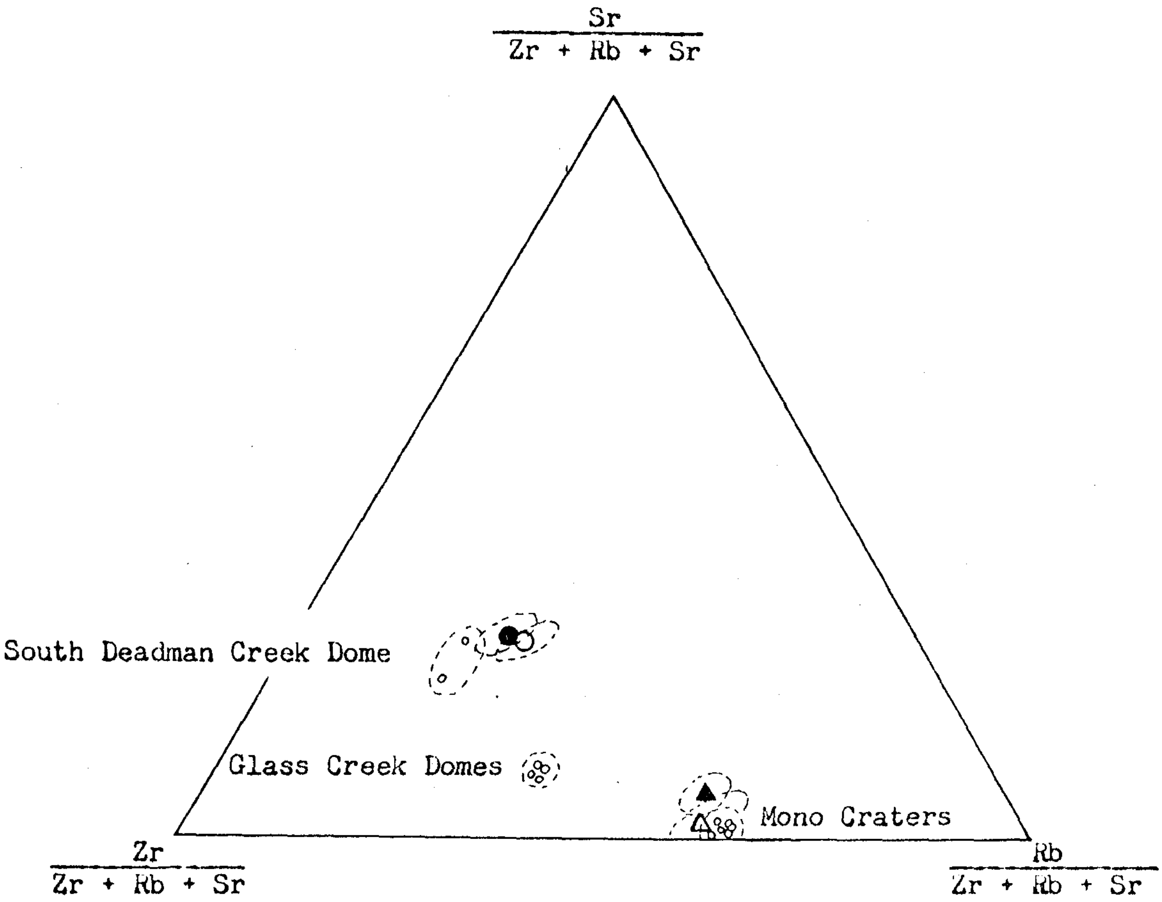
Plotting of the Sr, Rb, and Zr data on the ternary diagram (figure 7) published by Jack and Carmichael (1968) establish likely correlations to specific eruptive centers in the volcanic chain.

Tephra 1 appears to be related to an eruptive vent beneath the south Deadman Creek obsidian dome which partially covers a 1-kilometer tephra-ringed crater. The match is not perfect, for the analyses by Jack and Carmichael show some variability among the several samples from this vent; however, this is the only known vent with Sr levels on the order of 200 ppm.

Tephra 2 correlates best with the northern Mono Craters characteristic of very low Sr content. On the basis solely of trace-element chemistry, it is not possible to identify exactly which of the tephra-ringed craters of the northern group is the source of tephra 2. The 1 kilometer diameter tephra-ringed crater of Panum Butte or the unnamed crater just to the southwest of it are likely candidates. The rather massive tephra ring of Panum Crater contains relatively little coarse pumice lapilli. In a road cut this ring is seen to consist of stratified sheets of ashy fine lapilli, an observation consistent with the relatively fine texture of tephra 2 as seen in Lee Vining Canyon and in the Bodie Hills where its greater than 1-foot thickness indicates proximity to a source. Other vents with large tephra rings, having compatible geochemistry with tephra layer 2, may lie beneath the three large coulee flows of the main Mono Crater complex. Such hidden vents could just as well have served as the source of tephra layer 2. Careful mapping and establishment of relative temporal relationships of eruptive sequences within the geochemically homogeneous northern group of extrusives are required to establish with certainty which vent erupted the widespread, fine-textured tephra of layer 2.

Age of the Eruptions

Ages of tephras 1 and 2 are established by radiocarbon dates on well preserved conifer logs stratigraphically associated with the tephras. It is useful to date a known growth increment of wood and then count annual rings to the center and to the exterior rim



Tephra 1
- ● lapilli from San Joaquin drainage near Devil's Postpile National Monument.
- ○ volcanic ash from Sugarloaf Valley, Kings Canyon National Park

Tephra 2
- ▲ volcanic ash from the north fork of the Kings River drainage
- △ volcanic ash from East Meadow of Aspen Valley, Yosemite National Park.

Figure 7. Trace-element correlation diagram for tephra layers in the southern Sierra. Small open circles are analysis by Jack and Carmichael (1968) of rocks from the Mono and Inyo Craters. Corners of the ternary plot are concentrations in ppm normalized to the sum of concentrations of Zr, Rb, and Sr. Dashed lines are the estimated precision of analysis.

to date the establishment and death of the tree. Suess (1965) shows that there is no transfer of radiocarbon into the sapwood of earlier grown annual rings; therefore, radiocarbon activity of a known growth interval is unique (except for secular variations of radiocarbon in the atmosphere) and the radiocarbon age can be extended by annual rings through the lifetime of the tree.

Redating the Inyo Crater Lakes: Exposed in the walls and rim of the southern Inyo Crater Lake explosion pit is a heterogeneous mixture of mud and rock apparently erupted from the pit without any associated magma. The rim material overlies a 2-foot bed of pumice lapilli, identified as tephra 1. Incorporated into the rim material are uncharred, well-preserved, conifer logs. Rinehart and Huber (1965) obtained an age of 650 ±200 years B.P. upon one of these (W-1431). This much quoted age for the youngest expression of volcanism in the area is a maximum for the explosion that excavated southern Inyo Crater Lake. It also fixes a minimum age for tephra layer 1.

With the better counting precision now available in radiocarbon laboratories, it seemed worthwhile to try to reduce the uncertainty in the age of these logs. The log probably sampled by Rinehart and Huber has 185 annual rings which introduced a further uncertainty in the W-1431 age. Analyses of the 15 to 35 year old growth increment of heartwood yielded a radiocarbon age of 710 ±60 years B.P. (UGa-603). Counting annual rings both ways from the dated growth increment shows that this tree was established as a seedling at radiocarbon age 725 ±60 years B.P. and died at 550 ±60 years B.P. If the eruption at the southern Inyo Crater Lake site killed this tree (that is, it did not incorporate a dead log lying on the surface) then the death age of the tree virtually dates that eruption within 60 years on either side of 1400 A.D. (conversion to calendar age uses the curve of Suess, 1970).

Age of tephra 1 and the south Deadman Creek eruptive sequence: The south Deadman Creek eruptive vent which produced tephra 1 lies 1.5 km north of the Inyo Crater Lakes. It is almost certain that this pumice eruption would have killed, or at least charred, vegetation within several kilometers of the vent, especially in the direction of the Inyo Crater Lakes which lie along the axis of the main lobe of tephra distribution. The dated log and other logs in the rim material of the crater lake are uncharred (Rinehart and Huber, 1965) and therefore must have established in the area after eruption of tephra 1. Thus the date of 725 ±60 years B.P. indicates only a minimum age for the eruption of tephra 1.

On the west slope of the Sierra near the village of Dinkey Creek, a meadow soil sequence contains tephra layer 1 resting directly upon a fir log of 70 annual rings which apparently fell into the meadow when it died of natural causes. The wood 40 to 52 years older than the death age of the tree yields a radiocarbon age of 760 ±60 years B.P. (UGa-602) which corrected to the death age gives a maximum of 720 ±60 years B.P. for the eruption and deposition of tephra 1. This remarkable agreement of ages, obtained at separate sites, confirms the radiocarbon age of the pyroclastic eruption on Deadman Creek within 720 ±60 years B.P. Applying a small radiocarbon secular variation correction from Suess, 1970, Plate 1, the calendar age of the tephra 1 eruption is established within 60 years of 1240 A.D. It may well be within the capability of dendrochronologic techniques (Ferguson, 1970, and Stokes and Smiley, 1968) to date the eruption of tephra layer 1 and the Inyo Crater Lakes even more accurately.

A date of 1440 ±150 years B.P. (W-727) on charred wood from beneath pumice on Mammoth Mountain is quoted (Lajoie and Carmichael, 1967, Rinehart and Huber, 1965, Bailey and others, 1973b, and Jahnda, 1967) as the maximum age for the surficial pumice that blankets the Mammoth Lakes area. Charcoal from beneath this pumice was collected during the construction of the Mammoth Ski Lodge and dated by Berger and Libby (1966) at 920 ±80 years B.P. Since the pumice overlying these dated materials is tephra 1 which erupted 720 ±60 years B.P. these greater ages must have been obtained from material that had already been dead for several hundred years prior to burial by the pumice shower.

Age of tephra 2: The radiocarbon age of tephra 2 is established by the fortuitous occurrance of this fine ash directly upon a stump and beneath a log in meadow deposits near the North Fork of the Kings River. The outermost wood from the stump and log yielded radiocarbon ages 1210 ±55 (UGa-450) and 1175 ±65 years (UGa-451) B.P. respectively. Averaging the limiting ages establishes a radiocarbon age within 78 years of 1187 years B.P. which can be rounded to 1200 years B.P. Applying the variation correction to these ages (Suess, 1970) gives a calendar age within 80 years of 750 A.D.

If tephra 2 was erupted by the Panum Crater vent, the radiocarbon age is in reasonable agreement with the obsidian-hydration rind age of 1300 years obtained by Friedman (1968) on the plug that lies within the crater. This suggests that the other ages obtained by

Friedman (1968) on the coulee flows in the Mono Crater complex (figure 2) may also be relatively accurate.

A previously reported age of 1545 ±90 years B.P. for tephra 2 at a site in western Yosemite National Park (Wood, 1972) was based on charcoal derived from material that probably had already aged several hundred years before incorporation into the water-lain tephra layer.

An age of 700 ±200 years B.P. was obtained (Rubin and Alexander, 1960) from material reported to lie between two ash layers in a bank along Bodie Creek. The upper ash has a thickness continuity characteristic of the 1200 year B.P. tephra 2. These ash layers have not yet been trace-element analyzed; however, this sequence suggests the possibility of another young eruption, not detected in the Sierras, from which ash blew to the northeast. The material between the ash layers contains carbonized roots suggesting that 700 year B.P. age may suffer from younger carbon contamination.

Volume of erupted magma: Volume estimates (table 3) are based upon crude isopach data at localities shown on figures 4 and 5. It is not always possible to discriminate deposits secondarily thickened by slope wash or slumping and from true air-fall deposits. In all cases, a conservative judgment was made, and the estimate should be considered a minimum. Mass and volumes given in table 3 are adjusted for porosity, since fine ash deposits have densities of about 1.5 gm/cm^3, course lapilli beds about 0.5 gm/cm^3, and obsidian domes about 2.3 gm/cm^3. Reduced volumes are given in terms of the obsidian dome density. The depositional lobe of tephra 1 is well delineated and 0.2 cubic kilometers is a good estimate of the volume of erupted magma. Since tephra 2 is delineated only west of its source, and the main lobe may have formed in some other direction, its volume may be considerably more than the computed 0.2 kilometers.

Other dated tephra layers: Another radiocarbon age bearing on the eruptive history of the Mono-Inyo volcanoes was obtained from a section near the Devils Post Pile National Monument. Here, tephra layer 1 and 2 overlie two deeper volcanic ash layers (figure 8). Charcoal collected from a layer of charred material just beneath the third deepest layer yields a radiocarbon age of 3365 ±140 years B.P. These two deeper layers are also encountered in other sections in the upper San Joaquin River drainage, but they have not yet been adequately studied or analyzed for trace-elements, so the source vent remains unknown.

Figure 8. Tephra stratigraphy exposed in a dissected meadow flat in the valley of the upper middle fork of the San Joaquin River, by Starkweather Lake near Devils Post Pile. The meadow has not regrown upon the 720-year old pumice layer. Stadia rod gives scale of the foreground only. The top of the bank is 11 feet above tephra layer 4. The 3-foot long boulder has recently rolled into the gully.

Tephra 1 : a two foot layer of coarse pumice lapilli, maximum diameter = 4 cm, unstratified.

Tephra 2 : a 0.3-foot layer of pumiceous ash, maximum diameter = 0.5 mm, unstratified.

Tephra : a 0.5-foot layer of pumiceous ash, graded air-fall stratification, basal pieces up to 2mm diameter.

Radiocarbon age on charcoal layer = 3375 $\pm$ 140 B.P.

Tephra : a 0.3-foot layer of pumiceous ash, maximum diameter = 4.5 mm, unstratified.

TABLE 3. ERUPTION VOLUME ESTIMATE

	assumed density (gm/cm3)	*erupted weight-mass* (1015 gms)	*erupted volume* (reduced to= 2.3 gm/cm3 cubic kms
Tephra 1 (720 years B.P.)			
tephra farther than 8 km	1.5	0.16	0.07
tephra closer than 8 km	0.5	0.19	0.04
Deadman Creek south obsidian dome	2.3	0.12	0.06
Total tephra 1 magma	———————	0.38	0.17
Tephra 2 (1200 years B.P.)			
tephra farther than 3 km	1.5	0.48	0.21
tephra closer than 3 km	1.0	0.02	0.01
Panum Butte obsidian dome	2.3	0.02	0.01
Total tephra 2 magma	———————	0.52	0.23

Other sections with tephra layers are known in the vicinity of the Mono–Inyo Craters, in the White Mountains (Marchland, 1975), and in a dated lacustrine sequence beneath Black Lake described by Batchelder (1970) where 5 or possibly 6 tephras occur.

Conclusion

Known ages of eruptive centers within the chain are summerized in figure 2. Stratigraphic relationships and radiocarbon ages of tephra sequences (figure 9) summarize the known tephrachronology resulting from magmatic eruptions within the chain. This tephrachronology indicates activity spanning the last 40,000 years characterized by periods of dormancy 500 to a few thousand years in duration.

Tephra 1 (1250 A.D. ±60 years) and tephra 2 (750 A.D. ±80 years) erupted from opposite ends of the volcanic chain, appear to represent the most recent explosive eruptions of magma. The most recent dated expression of volcanism in the area is a phraetic explosion 1.5 miles south of the source vent of tephra 1 occurred between 1320 and 1460 A.D

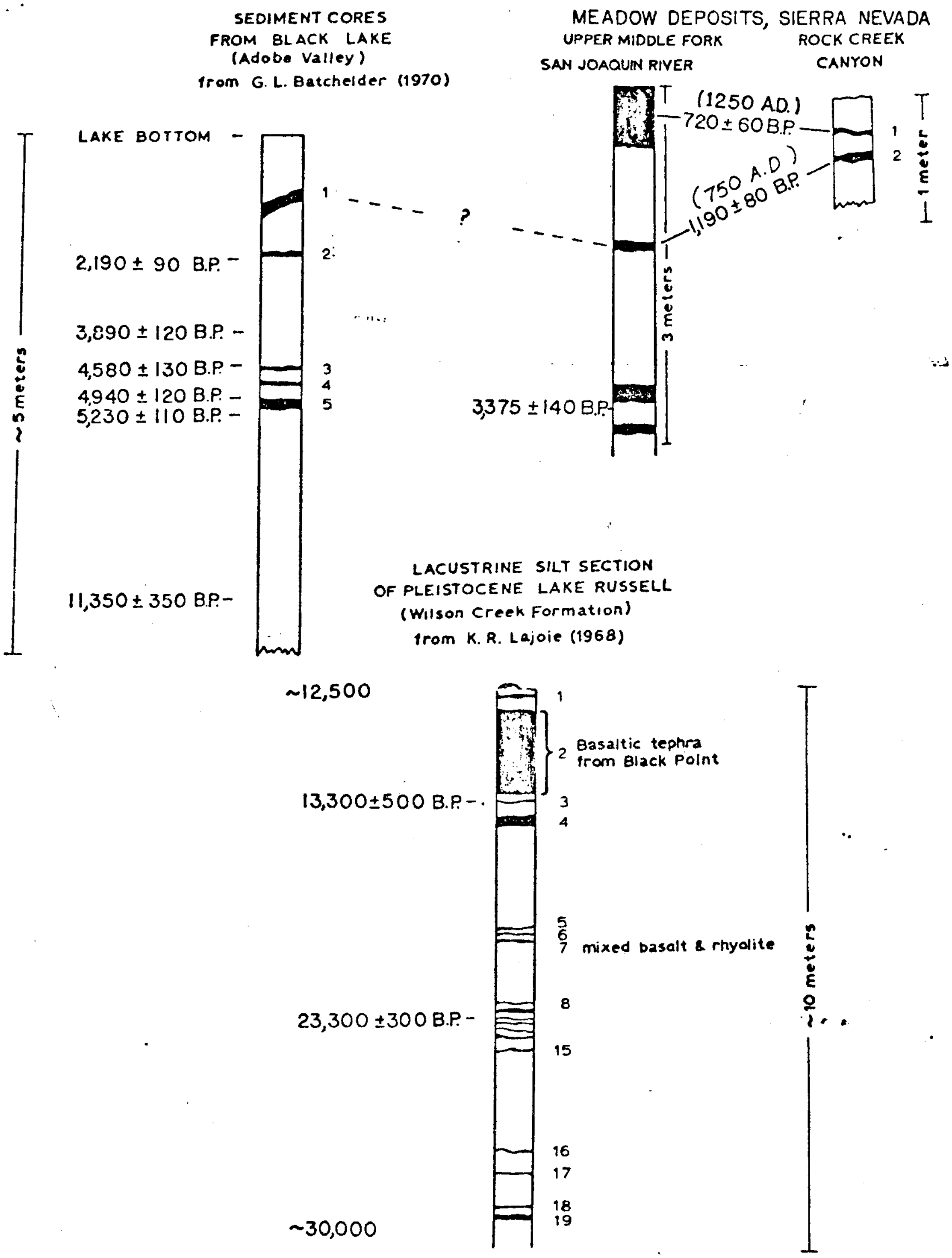


Figure 9. Summary of known tephrachronology for the Mono-Inyo crater. Tephras are rhyolitic to rhyodacitic except as noted. Thicknesses of tephra layers apply only to the particular locality.

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APPENDIX A --- INVENTORY OF MEADOWS DISSECTED BY GULLY EROSION

PARTIAL INVENTORY OF SIERRA NEVADA MEADOWS DAMAGED BY GULLY EROSION (includes only those meadows that have been inspected between 1970 and 1974)

Source of gully age
P = air photo
R = verbal report
T = tree-ring

LOCATION	15' QUADRANGLE	LOCATION	ELEVATION (feet)	AREA (acres)	AMOUNT LOST BY EROSION (acres)	SLOPE	DRAINAGE AREA (mi^2)	WIDTH (feet)	AGE OF GULLY MAX. (years A.D.)	AGE OF GULLY MIN. (years A.D.)
Stanislaus National Forest										
Sardine Meadow	Sonora Pass	Sec 1,T5N,R21E	8800							
Yosemite National Park										
East Meadow of Aspen Valley	Hetchy Hetchy	Sec 28,T2S,R20E	6170	15	15	0.04	2.3		1911(R)	1930(T)
Upper Tamarack Creek	Lake Eleanor	Sec 11,T2S,R20E								
Sierra National Forest										
California Creek	Bass Lake	SW¼,Sec 17,T6S,R22E							1944(P)	
Boggy Meadow	Bass Lake	NW¼,Sec 34,T6S,R22E	6400							
Beasore Meadow	Shuteye Peak	Sec 5,T6S,R23E	6800	83	3	0.03	2.8			1948(R)
Mugler Meadow	Shuteye Peak	Sec 22,T5S,R23E	6880	25	15	0.04	2.1			1945(R)
West Fork, Little Jackass Creek	Shuteye Peak	NW¼,Sec 15,T6S,R24E	5840	4	4	0.04	2.1			1944(P)
Upper Cabin Meadow on Laurel Creek	Huntington Lake	NW¼,Sec 14,T10S,R26E	7040	7	7	0.05	1.1		1940(P)	1940(P)
Lower Cabin Meadow	Huntington Lake	SE¼,Sec 15,T10S,R26E	6720							
Exchequer Meadow	Huntington Lake	SE¼,Sec 10,T10S,R26E	7200	18	2	0.04	0.7		1940(P)	1957(T)
Snow Corral Meadow	Huntington Lake	SW¼,Sec 30,T10S,R22E	7100							
Uppermost Long Meadow	Patterson Mtn.	NW¼,Sec 10,T11S,R27E				0.01	1.1			
West Fork Long Meadow Creek	Patterson Mtn.	SW¼,Sec 10,T11S,R27E		6	6				1940(P)	
Long Meadow, Chiquito Creek	Shuteye Peak	NW¼,Sec 18,T 6S,R24E	5040	12						
Kings Canyon National Park										
Sugarloaf Meadow	Triple Divide Peak	Sec 1,12, T14S,R30 E	7280	32	1	0.05	1.3			
Boggy Meadow	Triple Divide Peak	NW¼,Sec 8,T14S,R31 E	7200	6	3	0.08	0.5		1926(R)	1934(T)
Sequoia National Park										
Halstead Meadow	Giant Forest	Sec 13,T15S,R27E	6880	19	1	0.07	2.1			
Sequoia National Forest										
Weston Meadow	Giant Forest	Sec 12,T14S,R28E	6800	32	12	0.06				
Woodward Creek	Giant Forest	SE¼,Sec19, T14S,R29E	~7000				0.7			1966(R)
Fish Creek Meadow (Little Kern Basin)	Hockett Peak	SE¼,Sec10, T20S,R32E	6150	13	8	0.026	2.9		c.a.1961(P)	1966(T)
Greys Meadow (upper meadow)	~~Camp~~ Nelson	SE¼,Sec 4, T20S,R32E	6100	13	2	0.06	0.9			
Lower Peppermint Creek	Hockett Peak	SE¼ Sec36,T21S, R32E	4900	—	—					

APPENDIX B

WATER TABLE MODEL

Changes in water-table configuration arising from climatological control of water supply or from locally altered transpiration draught are evaluated by a two-dimensional steady-state ground-water model. Sites evatuated are on gently sloping valley fill for which the geometry and hydraulic conductivity are approximately known. It is assumed that the water-table configuration is determined entirely by flow above a certain level which may either be stream level at the foot of a seepage face or the level of impermeable bedrock. Ground water flow below this level is assumed to be negligeable.

From Dupuit-Forcheimer theory (van Schilfgaarde, 1957) it follows that the flux of seepage water across a vertical plane in a homogeneous permeable media can be expressed:

$$q_x = -Kh \frac{dh}{dx}$$

where K = hydraulic conductivity

h = height of the water table above stream level or the assumed surface of zero hydraulic head.

x = horizontal distance measured from the point of seepage outflow at the stream bank.

A mass balance at any point x along the slope gives the equation

$$-q_x = Q_i - q_t(L - x)$$

where

Q_i = supply of seepage water from the upslope perimeter of the meadow. (volume of water per second per linear distance unit of perimeter)

q_t = the distributed rate of transpiration draught on the ground water. (volume of water per second per unit area)

L = length of slope covered by transpiring vegetation.

Equating the two expressions for q_x and integrating between limits x = 0, x = x; and h = h_o, h = h yields an equation for the height of the water table:

$$h(x) = \left(2x[Q_i - q_t(L - x/2)]/K + h_o^2 \right)^{1/2}$$

where

h_o = height of the seepage face (bank of stream from which water seeps)

Parameters that determine the water table configuration, h(x), can be estimated at the East Meadow site. The model is used to estimate the nature of hydrologic changes necessary to transform a forested site into a wet sedge meadow. Prior to dissection, East meadow was supported largely by ground water seeping from the surrounding slopes to the valley fill, for no surface water flows into the meadow area in mid-summer. Presently, the main trunk gully collects water seeping out of the valley fill along the walls of the gully. The stream flow of 0.5 cubic feet per second in the lower end of the gully represents most of the seepage water that formerly supported the meadow. This flow is proportioned over the 3200-foot perimeter of the meadow to derive an upslope seepage flux (Q_i) of 1.56×10^{-4} cfs/linear foot of perimeter. It is assumed that the stream flowed through the middle of the 400-ft wide meadow, so that one side (L = 200 feet) is represented in the model (figure B-1).

The rate of transpiration draught is assumed to be 21 inches per 90-day growing season as suggested by measurements discussed in Chapter 2. Transpiration draught of mixed conifer forest vegetation is probably similar to that of the meadow. Soil moisture depletion by an isolated Sugar Pine from a cylinder of soil 61 feet in diameter amounts to 22.6 inches per growing season (Ziemer, 1968). Therefore 21 inches per 90 days ($q_t = 2.25 \times 10^{-7}$ cubic feet of water per second per square foot area) can be assumed for either forest or meadow cover.

The most uncertain parameter is the hydraulic conductivity. Values measured by the auger-hole method (Luthin, 1957) in the upper

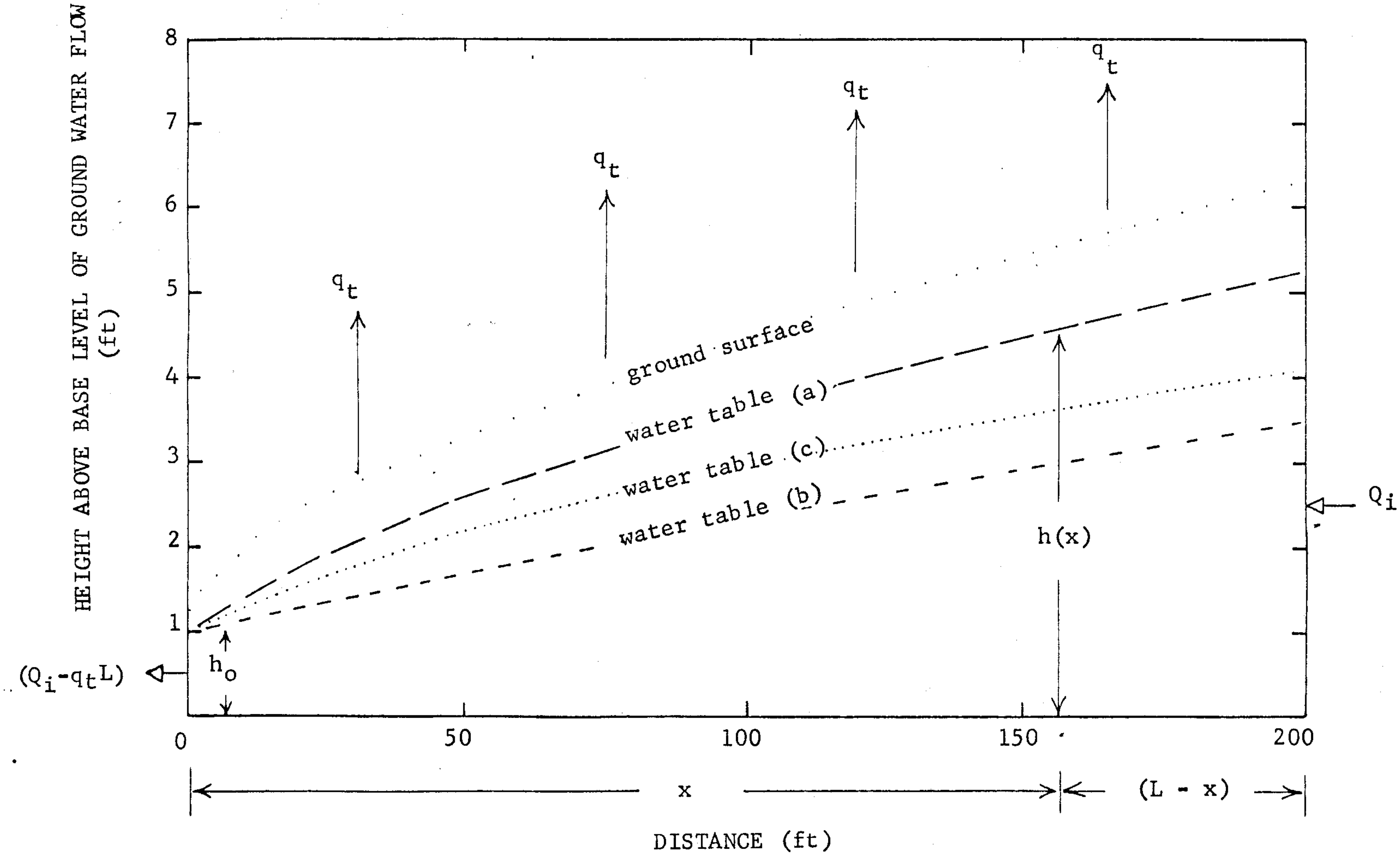

Figure B-1. Two-dimensional, steady-state, model of water table configuration in permeable valley fill deposits. Water table (a) simulates meadow conditions, water table (b) simulates forest conditions, and water table (c) simulates elimination of transpiration draught, q_t, by forest fire. Upslope seepage supply, Q_i, of case (a) is twice that of case (b) and (c).

5 feet of meadow deposits range 80 - 800 ft/day. Low values are in sandy loam and high values are caused by intercalcated well-sorted sand layers. The meadow surface slope of 2.1 per cent is an additional constraint on the model, for the water table must approximately parallel the meadow surface. A value of K equal to 170 feet per day (0.002 feet/second) gives a water-table slope of 2.1 per cent over a distance of 200 feet. Significantly different values for hydraulic conductivity give unreasonable slopes

Three cases are illustrated in figure B-1. Case(a) simulates meadow conditions using $K = 0.002$ ft/sec, $q_t = 2.25 \times 10^{-7}$ cfs/ft^2, and $Q_i = 1.56 \times 10^{-4}$ cfs/linear foot. The land surface is shown 1 foot above the meadow-supporting water table.

Prior to meadow conditions, a mixed-conifer forest thrived on the site (Chapter 2, page 52 - 57) indicating a water table several feet deeper. In order to simulate forest conditions, the upslope seepage supply must be diminished at least 50 per cent its value in case (a) or $Q_i = 0.78 \times 10^{-4}$ cfs/linear foot. This re-establishes the water table 1.2 to 1.7 feet deeper than in case (a), or at a total depth near 3 feet as shown in figure B-1, case (b). A 3-ft deep water table is marginal for a forest (figure 2-4, page 13).

Case (c) simulates the effect of a major forest fire which presumably kills all transpiring vegetation in the forested valley bottom of case (b). The upslope seepage supply remains the same, $q_i = 0.78 \times 10^{-4}$ cfs/linear foot, but the transpiration draught, q_t is set equal to zero. The water table rises 0.6 ft from its position in case (b), showing that this effect is small, and not a likely cause for the conversion of a forested valley bottom to a wet meadow.

APPENDIX C

COMPUTATION OF THE HYDRAULIC NATURE OF FLOOD FLOW ACROSS A MEADOW

A stage discharge relationship can be formulated from the Manning equation for uniform, turbulent flow in an open rough, rectangular channel.

Q = total discharge through channel (cubic feet per second)

A = cross-sectional area of channel

D = depth of flow

w = width of channel

r = hydraulic radius = $\frac{\text{cross-sectional area}}{\text{wetted perimeter}} = \frac{w\,D}{w + 2D}$

S = energy gradient of channel which is assumed to be the slope of the bed.

V = mean flow velocity

n = Manning roughness coefficient which has been computed from from measurements of mean velocity and channel geometry on similar channels (coefficient valid only for Manning equation computations in units of feet and seconds).

For a wide, rectangular, open channel:

$$A = w\,D$$

$$r = D \quad \text{for } w \gg D \tag{1}$$

The Manning equation is written:

$$V = \frac{1.49}{n}\, r^{2/3}\, S^{1/2} \tag{2}$$

Discharge through the channel can be written:

$$Q = w\,D\,V \tag{3}$$

From (1), (2), and (3):

$$Q = \frac{1.49}{n}\, D^{2/3}\, S^{1/2} \quad \text{and} \quad D = \frac{n\,Q}{1.49\,w\,S^{\frac{1}{2}}} \tag{4}$$

Computations using (4) must be carried out in foot - second units.

The most uncertain parameter in this stage-discharge relationship is the Manning roughness coefficient, n. Chow (1959) lists values of 0.03 to 0.05 for grass-lined channels with velocity-depth products greater than 3 ft^2/second (figure 7-14 in Chow, 1959), and 0.03 to 0.05 for mountain streams with gravel beds (table 5-6 in Chow, 1959). A value of n = 0.04 is assumed in Chapter 2, page 38, for over-bank flood flow across a meadow; however, uncertainty in the assumed value for n indicates that the computed flow depth may be in error by a factor of $(0.05/0.03)^{4/3} = 2$.

Competency of flow can be estimated by computing the shear stress on the bed using DuboYs equation (Blatt and others, 1972). The DuboYs equation for gravity-driven flow balances the component of the weight force tangential to the bed with the force of the shear stress on the bed(Chow, 1959).

$$\tau = g \rho_f S D \qquad (5)$$

where

$g\rho_f$ = specific weight of water.

S = channel energy gradient = slope of bed in uniform flow.

D = depth of flow

τ = shear stress on the bed

Using the depth computed in equation (4), shear stress on the bed can be computed with equation (5).

Experimental work by Shields relates the bed shear stress required to initiate motion of a particle on a flat bed (Vanoni, 1966, and Blatt and others, 1972). A simplified version of Shields diagram is provided by Blatt and others (1972) for initiation of of quartz sand on a plane bed (water temperature at 16°C). Using the bed shear stress from equation (5), the size of the largest particle set in motion by the flow is read from the diagram (figure C-1).

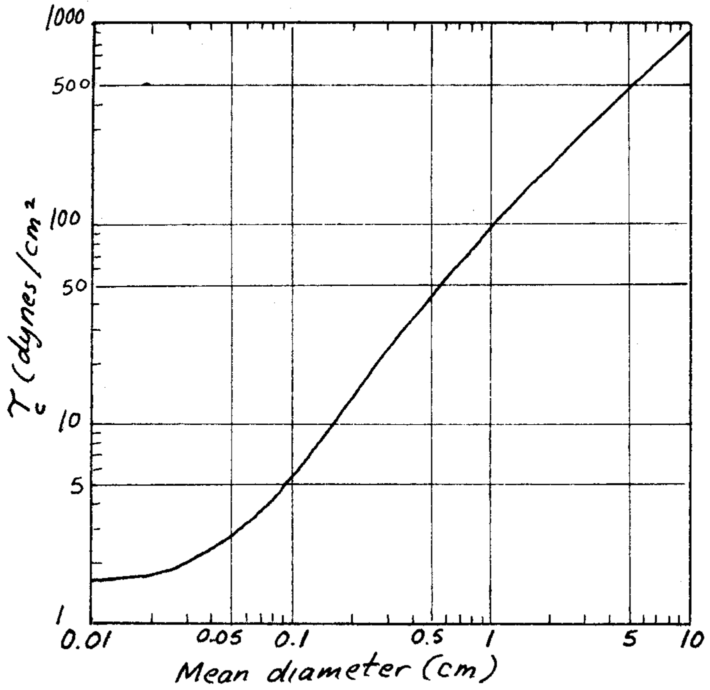

Figure C-1. Critical bottom shear stress (dynes/cm^2) for initiation of motion of quartz sand on a plane bed (from Blatt, Middleton, and Murray, 1972)

The procedure given here can also be reversed to estimate the flood discharge necessary to initiate motion of a given particle size.

APPENDIX D

[RADIOCARBON, VOL. 17, NO. 1, 1975, P. 99-111]

UNIVERSITY OF GEORGIA RADIOCARBON DATES IV

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The following list of dates is compiled from samples prepared since publication of our last date list (R, 1974, v 16, p 131-141). The counting equipment and operating procedures are the same. Ages are quoted with a 1σ counting error which includes statistical variation of the sample count as well as for background and standard, using AD 1950 as the reference year and 95% NBS oxalic acid for ^{14}C dating as the standard. The half-life value used is 5570 years.

3375 ± 140
UGa-449. Devil's Post Pile National Monument **1455 BC**

Charcoal, 2.8m deep in 3.7m sec of dark yellowish-brown loamy soil exposed in gully ca 20m W of rd to Reds Meadow, 0.3km S of Starkweather Lake (37° 39′ 32″ N, 119° 04′ 18″ W). Four volcanic ash layers occur in this sec. This dated charcoal layer immediately underlies 3rd deepest volcanic ash layer and dates maximum age.

West Fork of Long Meadow Creek series

West Fork site, Sierra Nevada (36° 58′ 47″ N, 119° 0′ 49″ W).

1210 ± 55
UGa-450. Wood **AD 740**

Outer wood from 0.5m diam, white fir stump *in situ* buried 0.6m beneath surface of a recently dissected meadow, 1.2h.

1175 ± 65
UGa-451. Wood **AD 785**

Outer wood from 0.5m diam white fir log buried 0.45m beneath meadow surface.

3320 ± 85
UGa-604. Wood **1370 BC**

Fir log, 0.2m diam buried 3.1m.

3770 ± 65
UGa-605. Wood **1820 BC**

Fir log, 0.1m diam buried 5.8m.

UGa-602. Heartwood from Lower Cabin Meadow, **760 ± 60**
Sierra Natl Forest **AD 1190**

Heartwood comprised of 12 growth rings 40 to 52 yr older than outermost growth ring of fir log, 0.4m diam, buried 0.3m and resting directly on rhyolitic tephra, 2cm thick.

UGa-621. Wood Upper Cabin Meadow **8960 ± 90**
on Laurel Creek **7010 BC**

Pine wood buried 3.05m in basal paleosol of Upper Cabin Meadow on Laurel Creek, Sierra Natl Forest.

UGa-622. Wood **9855 ± 100**
7905 BC

Heartwood from pine stump, 1m diam, buried 3.1m in Exchequer Meadow, Sierra Natl Forest.

East Meadow series

Wood and charcoal from several secs alluvial and organic fill exposed in gullies that dissected this 6.1h meadow, 2.5km E of Aspen Valley, Yosemite Natl Park, California (37° 50′ N, 119° 44′ 30″ W). Coll 1972 and subm by S H Wood.

UGa-447. Charcoal **9020 ± 270**
7070 BC

Charcoal from the "O" horizon of paleosol with O, A, C profile developed upon granitic bedrock. Paleosol was buried 7.5m beneath meadow surface.

UGa-452. Wood **9480 ± 90**
7530 BC

Wood from pine stump, 0.3m diam, *in situ*, rooted in mucky soil extending laterally to basal paleosol of UGa-447. Sample was buried 10m beneath meadow surface.

UGa-448. Wood **2830 ± 65**
880 BC

Wood from root mass *in situ* of fir stump, 1m diam, truncated and buried 2.5m beneath meadow surface.

General Comment: dates, with I-6049 (unpub) establish chronology for 3 depositional units that comprise sub-meadow stratigraphy described by Wood (1973).

I-6049 - C^{14} = 175 ± 9 1545 ± 90 B.P.
A.D. 405

Charcoal lumps in a 3-cm thick layer of volcanic ash 1 foot deep in section exposed by gully at East Meadow, 1.5 miles east of Aspen Valley, Yosemite National Park, California. (37°50'N, 119°44½'E).

Boggy Meadow site series

Kings Canyon National Park (36° 43′ 30″ N, 119° 30′ W).

UGa-620. Wood **8705 ± 95**
6755 BC

Wood from stump, 0.07m diam, rooted in basal paleosol, buried 6.1m in Boggy Meadow.

UGa-623. Charcoal **10,185 ± 105**
8235 BC

Pieces of organic soil containing charcoal buried 6.7m in Boggy Meadow, Sugarloaf Valley.

UGa-603. Pine log **710 ± 60**
AD 1240

The 20-yr growth ring interval 150 to 170 yr older than outermost growth ring of pine log, 0.2m diam, buried in a volcanic mudflow, exposed on N wall of S-most explosion pit of Inyo Crater Lakes, Inyo Natl Forest, California.

UGa-674 Beasore Meadow, Sierra National Forest 7705 ± 90 B.P.
5755 B.C.

The annual ring interval from age 31 to 47 from a 56-year longevity fir log buried 12.5 feet beneath the meadow surface.

UGa-675 Tuolumne Meadows, Yosemite National Park 3930 ± 80 B.P.
1980 B.C.

Charcoal lumps from a charcoal-rich layer in fluvial gravels 3.6 ft beneath meadow surface.

APPENDIX (CHAPTER 4) -- TRACE-ELEMENT ANALYSIS

The procedure used to obtain clean glass separates from volcanic ash samples contaminated with mineral silt and sand grains, and the rapid-scan method of X-ray fluorescence analyses for these trace-elements, are taken from Sarna-Wojcicki (1971, and personal communication, 1973). The reader is referred to his thesis for more complete details. The particular procedures for the results reported herein are as follows:

1) Raw volcanic ash samples or crushed lapilli grains are wet-sieved in acid-cleaned (conc. HNO_3) stainless steel and polyethylene screens, the 80 to 120 mesh fraction retained for analysis.

2) By varying field intensity, slope, and tilt of the Franz magnetic separator it is possible to separate first the magnetic mineral grains from the salic mineral and glass grains, and second, to separate the glass grains from the salic mineral grains. Volcanic glass generally has a relatively high iron content and hence a magnetic susceptability significantly greater than the salic mineral grains, but significantly less than most mafic minerals. Several runs through the separator may be necessary to obtain sufficiently pure glass during the course of which some glass is lost in the salic mineral separate. The glass separate is checked for purity by microscopic examination of grain mounts under crossed nichols.

3) Glassy separates are washed in distilled water and dried. Strontium carbonate precipitation and other contamination problems of soil environments did not pose a problem with these samples from acidic mountain soils. Samples from semi-arid soils and lacustrine sediments often need to be subjected to short HCl acid wash and an HF acid etch.

4) About 5 grams of weighed sample are mixed with standard grade Whatman cellulose powder or other binder material in a 4:1 ratio and hydraulically pressed into wafers for the sample holders of the X-ray fluorescence unit. Standard samples are prepared in a similar manner.

5) The samples are irradiated with X-rays from a W-tube operated at 50 kilovolts. The X-ray fluorescence intensity counts are converted to concentrations in parts per million by counting the intensities of similarly prepared samples of U.S.G.S. rock standards W-1, G-1 and a mixture of spectrographic grade quartz and G-1 both before and after each unknown sample. Concentrations of Sr, Rb, and Zr are then read from the calibration curve prepared by plotting the data from the standards.