

INFLUENCES OF QUATERNARY CLIMATIC CHANGES ON PROCESSES OF SOIL DEVELOPMENT ON DESERT LOESS DEPOSITS OF THE CIMA VOLCANIC FIELD, CALIFORNIA

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SUMMARY

Soils formed in loess are evidence of both relict and buried landscapes developed on Pliocene-to-latest Pleistocene basalt flows of the Cima volcanic field in the eastern Mojave Desert, California. The characteristics of these soils change systematically and as functions of the age and surface morphology of the lava flow. Four distinct phases of soil development are recognized: phase 1 – weakly developed soils on flows less than 0.18 M.y. old; phase 2 – strongly developed soils with thick argillic horizons on 0.18 - 0.7 M.y. old flows; phase 3 – strongly developed soils with truncated argillic horizons massively impregnated by carbonate on 0.7 to 1.1 M.y. old flows; and phase 4 – degraded soils with petrocalcic rubble on Pliocene flows. A critical aspect of the development of stage 1 soils is the evolution of a vesicular A horizon which profoundly affects the infiltration characteristics of the loess parent materials. Laboratory studies show that secondary gypsum and possibly other salt accumulation probably occurred during the period of phase 1 soil development. Slight reddening of the interiors of peds from vesicular-A horizons of phase 1 soils and presence of weakly developed B horizons indicates a slight degree of in situ chemical alteration. However, clay and Fe oxide contents of these soils show that these constituents, as well as carbonates and soluble salts, are incorporated as eolian dust. In contrast to phase 1 soils, chemical and mineralogical analysis of argillic horizons of phase 2 soils indicate proportionally greater degrees of in-situ chemical alteration. These data, the abundant clay films, and the strong reddening in the thick argillic horizons suggest that phase 2 and phase 3 soils formed during long periods of time and periodically were subjected to leaching regimes more intense than those that now exist. Flow-age data and soil-stratigraphic evidence also indicate that several major loess-deposition events occurred during the past ~ 1.0 M.y. Loess events are attributed to past changes in climate, such as the Pleistocene-to-Holocene climatic change, that periodically caused regional desiccation of pluvial lakes, reduction of vegetational density, and exposure of loose, unconsolidated fine materials. During times of warmer interglacial climates, precipitation infiltrates to shallower depths than during glacial periods. Extensive, saline playas which developed in the Mojave Desert during the Holocene are a likely source of much of the carbonates and soluble salts that are accumulating at shallow depths both in phase 1 soils and in the formerly noncalcareous, nongypsiferous argillic horizons of phase 2 and 3 soils.

1. INTRODUCTION

In the past decade eolian activity has been recognized as an important factor that strongly influences geomorphic and pedologic processes in deserts. Desert dust potentially affects climate by increasing backscatter of radiation to space (PEWE 1981). Desert dust is also a major source of secondary carbonates (LATTMAN 1983, GILE 1975, 1977, BACHMAN & MACHETTE 1977, MCFADDEN & TINSLEY 1985) and clay (YAALON & GANOR 1973,

GILE et al. 1981) in desert soils and of Fe and Mn oxides in desert varnish (ALLEN 1978, DORN & OBERLANDER 1981). Many geomorphic surfaces in desert landscapes, including rocky hillslopes (WELLS et al. 1982) alluvial piedmonts (GILE & GROSSMAN 1979), and lava flows (GREELEY & IVERSON 1981) efficiently entrap eolian sediment. DOHRENWEND et al. (1984) recently reported that extensive portions of the lava-flow surfaces of the Cima volcanic field in the Mojave Desert, California (Fig. 1) are buried by eolian

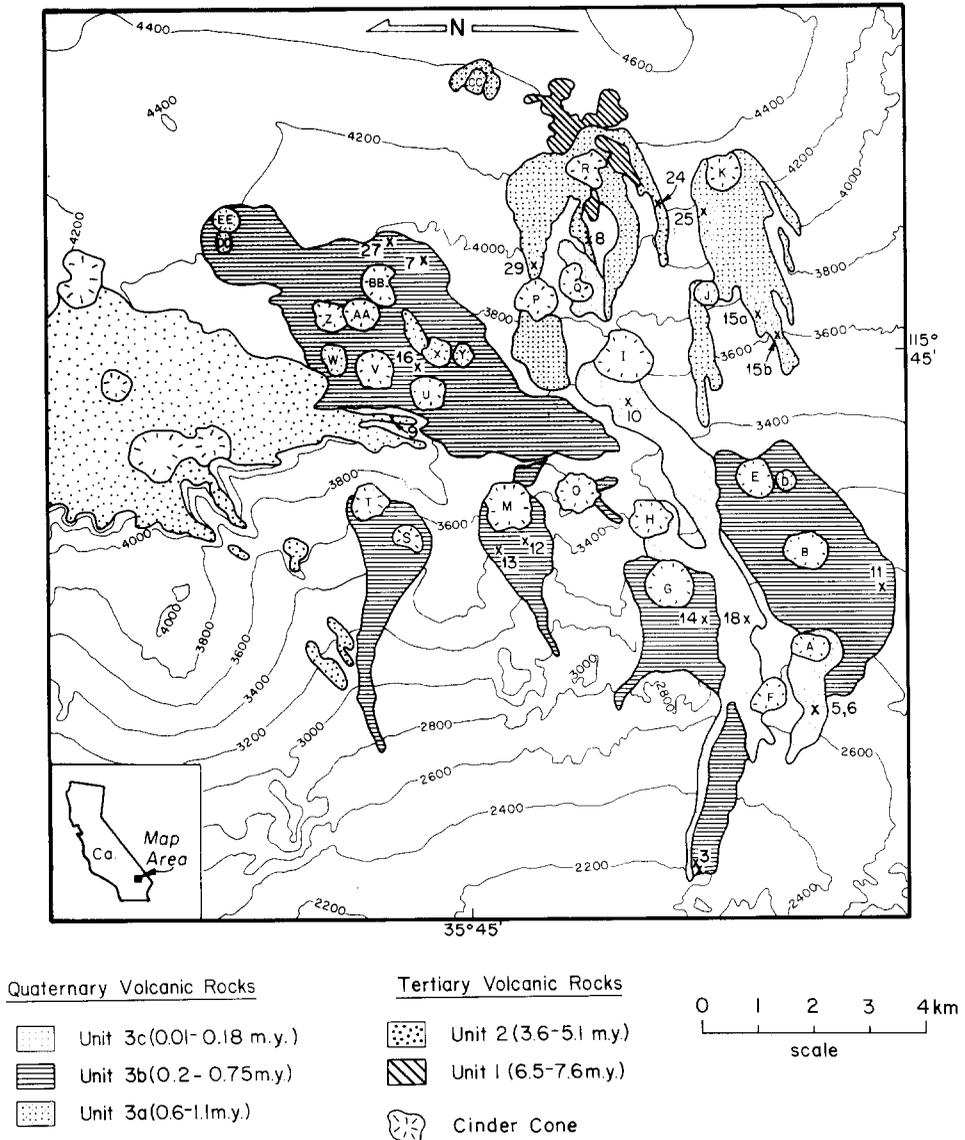


Fig. 1: Generalized map of the Cima volcanic field showing the location of major cones and flows and sites of soil profiles (given with prefix MP in Table 1) discussed in this paper. Volcanic vent, flow units, and cinder cone designations (capital letters) after DOHRENWEND et al. (1984). Elevation in feet.



Photo 1: Aerial oblique photograph of late and middle Pleistocene flows in the Cima volcanic field. Note constructional flow forms with eolian deposits in flow depressions on the younger, late Pleistocene flow and areas of stone pavements on both flows (arrows).

deposits in which soils have formed. These deposits are extremely well sorted, sandy silts which have been transported to flow surfaces predominantly as suspended load (WELLS et al. 1985), and thus we consider these deposits to be desert loess. Numerous studies have considered formation of soils in the extensive loess deposits of the North American mid-continent (RUHE 1983); however, there are no studies of formation of soils in desert loess. The Cima volcanic field provides an excellent setting for such a study. Sixty spatially distinct lava flows range in age from late Miocene to latest Pleistocene (DOHRENWEND et al. 1984, TURRIN et al. in press) and exhibit similar constructional morphology. In this paper, we describe the soil stratigraphy, soil-geomorphic relations, and major processes of soil development in loess deposits on the Cima flows. We also (1) propose a conceptual model for an evolutionary sequence of soil development that is based on the flow-emplacement chronology at Cima and (2) postulate impacts of Quaternary climatic changes on eolian activity and concomitant effects of climatic changes on geomorphic processes and pedogenesis that operate on these basalt flows.

2. STUDY AREA

The Cima volcanic field is located in the eastern Mojave Desert of California (Fig. 1) and contains approximately 40 basaltic cones and more than 60 associated lava flows that cover pedimented pre-Tertiary crystalline rocks and Tertiary gravel deposits (DOHRENWEND et al. 1984). Elongate flows in the volcanic field are thin (2.5 to 4.0 m thick), have relatively low gradients (2° to 5°), and exhibit surficial constructional forms that are characteristic of relatively fluid basaltic lavas (e.g. leveed flow channels, rafted spines and spires, and lobate to digitate flow margins) (Photo 1).

The climate of the central Mojave Desert at elevations between 600 and 1000 m is arid to semiarid (12 - 25 cm mean annual precipitation) and hot (16-18°C mean annual temperature) (NATIONAL OCEANIC AND ATMOSPHERIC ADMINISTRATION 1978). The present climate of the Mojave Desert is primarily the result of the rain-shadow effect of the high-altitude Transverse Ranges and the Sierra Nevada, located 100 to 200 km to the west and southwest. Much of the precipitation falls during relatively cool winters; hence, the effective soil moisture, as indicated by the leaching index of ARKLEY (1963), is as much as 2 to 4 cm at elevations over 1000 m (MCFADDEN 1982). Vegetation in the study area is classified as the Mojave Desert scrub community (BROWN et al. 1980). Typical xerophytic plants include cacti and small woody shrubs, such as creosote (*Larrea tridentata*), brittle bush (*Eucelia farinosa*) and Mormon tea (*Ephedra trifurca*). At elevations above 1300 m, Joshua tree (*Yucca brevifolia*) is common.

3. METHODS

3.1. FIELD METHODS

Soil profiles were described in pits excavated on the most level, geomorphically stable parts of K-ar-dated lava-flow surfaces (Fig. 1). These areas are typically characterized by interlocking, darkly varnished stone pavements. The soils were described according to methods and nomenclature of the SOIL SURVEY STAFF (1951, 1975). Pedogenic-carbonate content was estimated by determining morphological stages of carbonate accu-

mulation after GILE et al. (1966). Bulk samples were collected from all subhorizons; peds were collected from selected subhorizons for optical examination of micromorphology.

3.2. METHODS FOR QUANTIFICATION OF SOIL MORPHOLOGY AND TEXTURE

The degree of development of many soil properties is strongly time dependent (BOCKHEIM 1979). Indices combining horizon thickness and several soil properties have been defined that permit quantification of soil development (BILZI & CIOLKOSZ 1977, HARDEN 1982). The index of HARDEN (1982) is very strongly time dependent and increases at a relatively similar rate in diverse climatic settings and in differing parent materials (HARDEN & TAYLOR 1983, PONTI 1985, COLMAN & PIERCE 1986). We have calculated this index for soils on flows in the Cima volcanic field in order to compare the degree of soil development on different flows of many ages and to compare the degree of soil development in eolian deposits in the study area with the degree of development of dated soils formed in fluvial deposits from other chronosequences.

The Harden index is calculated by quantifying the difference ($x_{\Delta p}$) between values of soil properties (e.g. color, structure, dry consistence) and parent material values. A normalized index for each property is calculated by dividing $x_{\Delta p}$ by the maximum observed value for each property (x_{pmax}). Normalized values are summed for each property, divided by the total number of properties, and multiplied by the horizon thickness which yields the horizon products. The sum of the horizon products is the profile index, which, when divided by total profile thickness, determines the thickness-normalized index. The latter value is useful for evaluating differences in the degree of soil development that are not primarily due to soil thickness. Indices reported in this paper were calculated on the basis of six properties, including color, texture, structure, clay films, dry consistence, and secondary carbonate morphology.

3.3. LABORATORY METHODS

Soil samples were air dried and then subdivided into fine (< 2 mm) and coarse (> 2 mm) fractions. Sample splits of the fine fraction were made for determination of pH, electroconductivity (EC), particle-size distribution, and Fe-oxide composition. Soil pH was determined with a 1:10 soil-to-water ratio. Soil electroconductivity (EC), was measured in order to evaluate trends in soluble salt accumulation and was determined with a 1:5 soil-to-water ratio using a Sybron/Barnstead PM-70CB conductivity bridge. Soil particle-size distribution was determined after carbonates and soluble salts were removed by sodium acetate digestion. After clay dispersal by sodium phyrophosphate, wet-sieve analysis was used to determine the sand content and pipette extraction was used to determine the clay and silt content. A split of the fine fraction was crushed to a minimum of 80 mesh for determination of Fe oxyhydroxide content.

Wet-chemical extraction techniques were used to determine the content, composition, and relative degree of crystallinity of pedogenic Fe oxyhydroxides. The dithionite-citrate-bicarbonate method of MEHRA & JACKSON (1960) was used to determine total secondary (free) Fe oxyhydroxides (Fe_2O_3d). The oxalate extraction procedure of SCHWERTMANN (1964) and McKEAGUE & DAY (1966) was used to extract poorly crystalline, fine-grained Fe oxyhydroxides (Fe_2O_3o), mostly ferrihydrite (SCHWERTMANN & TAYLOR 1977).

Magnetite is removed from samples prior to Fe_2O_3 extraction due to magnetite solubility under the conditions of this procedure (RHOTON et al. 1981, WALKER 1983). Extracted Fe^{+3} was measured using a Perkins-Elmer 303 atomic absorption spectrophotometer. Total ferrous iron (FeOt), ferric (Fe_2O_3) and total ferric + ferrous iron ($\text{Fe}_2\text{O}_3\text{t}$) were determined for selected soils using hydrofluoric and sulfuric acid digestion followed by potassium dichromate titration to measure FeOt (KOLTHOFF & SANDELL 1961) and hydrofluoric + nitric + perchloric acid digestion to measure $\text{Fe}_2\text{O}_3\text{t}$ (HUSLER 1969).

Relative abundances of minerals in the $< 2\mu\text{m}$ fraction were determined for selected soils using X-ray diffraction techniques. A Phillips XRG-3000 diffractometer (Cu $K\alpha$ beam) was used to x-ray Mg-saturated, Mg-glycolated, K-saturated and heat treated (500°C) samples mounted on glass slides. Relative abundance of clay-mineral components was estimated on the basis of relative peak-height difference and changes in peak heights observed on diffractograms after different treatments (GRIM 1968).

4. SOIL MORPHOLOGY AND FLOW AGE

4.1. PHASE 1 SOILS

Soils formed in loess mantles on at least 17 different basalt flows that range in age from Pliocene to late Quaternary can be grouped into four major phases of soil development which are defined on the basis of morphological, textural, chemical, and mineralogical differences.

Only weakly developed soils are observed in loess deposits on all lava flows younger than 0.14 ± 0.04 m.y. Similarly developed soils also are present in thin deposits that overlie the more well-developed soils present on most older flows (Table 1). These soils, designated as phase 1 soils, exhibit a strongly developed vesicular A horizon (Av), a thin, reddened B horizon and a Stage I or Stage I + calcic horizon (carbonate stage morphology after GILE et al. 1966, and BACHMAN & MACHETTE 1977). A stone pavement forms the surface layer of this soil and consists of varnished basaltic clasts derived from flow rock and vent debris (Photo 1). Mean clast sizes in these pavements ranges between 3.3 and 10.0 cm, and clast relief above the soil surface ranges from 14 to 23 cm (WELLS et al. 1985). The surface of the stones is varnished; DORN (1984) has analyzed varnish from lava flow surfaces at Cima and determined that they resemble varnish from other regions of the Mojave Desert with respect to chemistry, mineralogy, and micromorphology. The stone bottoms that project into the vesicular horizon are commonly reddened to a hue of 5YR to 2.5 YR, attributable to accumulation of secondary Fe oxides.

The most clay and typically silt-enriched horizon of phase 1 soils is the vesicular A (Av) horizon (Tables 1, 2). Av horizons always exhibit strong, coarse platy to columnar structure (Photo 2). The spherical vesicles of Av horizons are attributed to entrapped soil air that expands as the soil temperature rapidly increases after summer rainfall events (EVENARI et al. 1974). In the study area these vesicles have thin, continuous coatings of calcareous clay and often become increasingly lenticular, subparallel to the soil surface, near the base of the horizon. The Av horizon is usually overlain by a very thin (0.1 to 0.5 mm) horizon (A) that is always less calcareous but is coarser-grained than the Av horizon. The A horizon lacks vesicles and its dry consistence is loose. In contrast, dry consistence of the Av horizon is slightly hard to hard. The outer several millimeters of Av horizon peds are also much less calcareous than the interior of the ped, which has disseminated calcium carbonate. Segregated carbonate also is present at the base of many Av peds. The interior of the ped also is typically redder than the exterior of the ped.

A thin, calcareous, and slightly reddened B horizon in which small amounts of clay and silt have accumulated usually is present below the Av horizon (Tables 1 and 2). Skeletal grains possess colloidal stains, and the uppermost part of the B horizon may exhibit weak vesicular pores. Pedogenic carbonate is present in this horizon and all underlying soil horizons. Carbonate is disseminated or is segregated as thin (0.1 to 1 mm) discontinuous coatings that occur mostly on the sides and bottoms of increasingly abundant basalt stones at depths below 10-15 cm. Occasionally, the carbonate coatings on the base of large stones are mantled by euhedral needles of secondary gypsum. The basalt stones, however, exhibit a minimal degree of physical weathering, sharply ringing to the blow of a hammer and possessing no obvious weathering rinds. The maximum depth of carbonate or soluble salt accumulation could not be determined because of the extreme difficulty of pit excavation in the increasingly basalt clast-rich soil. The lowest, clast-rich layer is composed of colluvial material and disaggregated flow rock and is called the rubble zone (WELLS et al. 1985). Figure 2 is a schematic diagram depicting the morphology of a typical phase 1 soil.

Textural analysis of the soil parent materials of a phase 1 soil on the youngest flow in the Cima volcanic field (WELLS et al. 1985) shows that the loessal parent materials are extremely well sorted, similar to well-sorted, though finer-grained desert dust analyzed by PEWE (1981). Quartz is not present in the basanitic basalts but is abundant in the silt and clay fractions of these soils, an observation also showing that the deposits are eolian. Petrographic studies of the parent materials show that trace amounts of basaltic fragments also are present in the parent materials. The basaltic material may be eolian, derived by deflation of fluvial deposits in the Cima volcanic field, or it may be derived by physical weathering of basalt on the flow surface. Values of the Harden index for phase 1 soils range between 11.90 and 17.66 (Table 1). These values fall within the range of index values reported by HARDEN & TAYLOR (1983) for soils formed on deposits of Holocene and latest Pleistocene age and indicates that the loessal parent materials of phase 1 soil are of this age. Such an age estimate is compatible with the minimum age estimate of the youngest flow in the Cima volcanic field on which a phase 1 soil is present.

4.2. PHASE 2 SOILS

The soils on flows between 0.14 ± 0.04 m.y. and 0.70 ± 0.06 m.y. old exhibit argillic B horizons as well as Stage II, II+, and rarely, Stage III calcic horizons. Moderately developed soils with these properties are designated phase 2 soils (Tables 1 and 2, Photo 2), and are commonly buried by an eolian deposit in which a phase 1 soil has formed. Often, the contact between phase 1 and buried phase 2 soils appears to be transitional over a 2-5 cm depth interval, although the contact occasionally is quite abrupt. The argillic horizon, identified on the basis of the presence of large quantities of translocated clay, ranges from 0.2 m to more than 1 m thick and occasionally extends downward into the rubble zone on the subjacent flow. The argillic horizon is typically strong brown (7.5YR 5/4 to 7.5YR 5/6), but may exhibit redder colors (5YR 5/4) that are associated with clay films. Soil structure ranges from granular to strong angular blocky. Significant clay has accumulated in the argillic horizon; however, clay content may vary significantly with depth rather than occurring as a well defined, single bulge typical of many argillic horizons (SOIL SURVEY STAFF 1975) (Tab. 2).

Secondary pedogenic carbonate exhibits primarily Stage II morphology and is present mostly as segregated material coating basaltic clasts or occurring as nodules and filaments.

Tab. 1: SUMMARY OF MORPHOLOGICAL DATA FOR SELECTED SOILS ON EOLIAN DEPOSITS ON FLOWS OF THE CIMA VOLCANIC FIELD; see Fig. 1 for location

Soil Profile: Maximum Phase	Flow Age ¹ (m.y.)	Horizon ²	Depth cm	Color (dry matrix)	Structure ²	Clay Films ²	Carbonate ² Morphology	Profile ³ Index	Thickness- Index ³
MP-5; 1	0.016 ⁴	A	0-0.2	10YR 7/3	sg	n.o. ⁵	n.o.	7.85	0.16
		Avk	0.2-4	10YR 7/3	3vcp/2mcpr	4np	m2ss		
		Bk	4-12	10YR 8/4	m & lfsbk	n.o.	d		
		2Bk	12-45+	10YR 7/3	m	n.o.	s		
MP-6; 1	0.016 ⁴	A	0-0.2	10YR 7/2	sg	n.o.	n.o.	7.85	0.16
		Avk	0.2-3	10YR 7/3	3vcp/2mcpr	4np	m2ss		
		Btk	3-7	7.5YR 6/4	2csbk	lnpf & co	s		
		Bk	7-45+	8.75YR 7/4	m	n.o.	s		
MP-18; 1	0.06 ± 0.03	A	0-2	10YR 7/4 & 8.75YR 7/3	sg	n.o.	n.o.	12.84	0.14
		Avk	2-6	10YR 7/4	3vcp/3mcpr	4np	m2ss		
		Bk1	6-32	10YR 4/6	m/1msbk	n.o.	csf		
		Bk2	32-69	10YR 4/4	m	n.o.	d		
		2Bky	69-85+	10YR 6/3	m	n.o.	d		
MP-10; 1	0.14 ± 0.04	A	0-1	10YR 7/3	sg	n.o.	n.o.	13.62	0.23
		Avk	1-2.5	10YR 7/3	3vcp/2mcpr	4np	m2ss		
		Btk	2.5-5.5	7.5YR 6/4	1nsbk	3npf & co	d		
		Bk	5.5-16	8.75YR 7/4	1msbk	n.o.	d, msf		
		Bkyl	16-60	10YR 6/4	1msbk	n.o.	d,s		
		Bky2	60+	10YR 6/4	m	n.o.	d,s		
MP-13; 1	0.39 ± 0.08 ⁶	A	0-0.3	10YR 7/3	sg	n.o.	n.o.	13.62	0.23
		Avk	0.3-2	10YR 7/3	3vcp/2mcpr	4np	m2ss		
		Btk	2-7	8.75YR 5/4	3vcp/vcsbk	3np	d		
		Bk1	7-26	10YR 7/4	m	n.o.	s,d		
		Bk2	26-70+	10YR 7/3	sg	n.o.	s,d		
MP-3; 2	0.33 ± 0.05 to ⁷ 0.58 ± 0.16	Avk	0-5	10YR 7/3	3vcp/vcsbk	4np	d, m2ss	13.62	0.23
		Btkb	5-19	5YR 5/6	2msbk/copl	2np, co	d		
		Bkb	19-28	7.5YR 6/4	m	co	s		
		Bkyb1	28-53	7.5YR 5/6	m	n.o.	s		
		Bkyb2	53-78	7.5YR 5/4	m/sbk	n.o.	d, 2sf		
		Bky3	78-90+	10YR 7/3	m	n.o.	d,s		

Tab. 1: continued

MP-16; 2	0.26 ± .05	Av Bwk Bky Btkyb1 Btkyb2 2Btkyb3 2Bkb1 2Bkb2	0-2 2-5 5-19 19-31 31-54 54-85 85-104 104-110+	10YR 7/3 7.5YR 5/4 10YR 7/3 7.5YR 6/6 7.5YR 6/4 7.5YR 6/6 10YR 7/6 10YR 7/5	3vcpl/mcpr 3mpl m 3f & msbk 3msbk & abk m/fsbk m/sbk m	4np0 3np0,co 1np0 4npf,co 2npf,co 1npf,co n.o. n.o.	d d lsc,d d,2sf d,2sf d,2sf d,s d,s	
MP-7; 2	0.27 ± 11	A Avk Bt Bk Btkb1 Btkb2 2Btkb3	0-0.2 0.2-7 7-10 10-22 22-37 37-48 48-81+	10YR 7/3 10YR 7/3 7.5YR 5/6 10YR 7/4 7.5YR 5/4 7.5YR 5/4 7.5YR 6/4	sg 3vcpl/2mcpr m/lmsbk m/lmgr 3fsbk 3fsbk m	n.o. 4np0 co n.o. 2npf&co 1npf&co co	n.o. m2ss d d,s ss ss d,ss,sc	
MP-27; 2	0.27 ± 11	A Avk ABvk Btk1 Btk2 2Btkyb 2Bkb	0-0.1 0.1-6 6-7 7-13 13-22 42-62 62-72+	10YR 7/3 10YR 7/3 7.5YR 6/4 5YR 5/6 7.5YR 6/6 6.25YR 7/4 10YR 7/4	sg 3vcpl/2mcpr 2mgr 3fabk/gr 3fabk/gr m/3mgr m/3mgr	n.o. 4np0 2npf & co 4npf & co 4npf & co 3npf & co co	n.o. d d,cssm d,frm mgm & sf d,msm d,cssm	34.88 0.47
MP-15; 2	0.33 ± 0.03	Avk BAvk Ckox 2Btkb1 2Btkb2 2Btkb3 2Bkb	0-5 5-7 7-24 24-34 34-54 54-74 74-90+	10YR 6/3 6.25YR 4/6 10YR 6/4 7.5YR 4/6 6.25YR 4/6 7.5YR 6/4 7.5YR 7/4	3vcpl/2mcpr 1mgr m 3f & msbk 3mabk 3mabk m	4np0 1npf & co n.o. 3npf & co 3npf&co,br 3npf,co,br n.o.	d d, clsf d,s d,c & msm d,msm d,cssm d,s	
MP-1; 2	0.33 ± .05 to 0.58 ± 0.16	A Avk ABk Btkyb1 Btkyb2 Btkyb3 2Btkyb 2Bkyb2	0-0.5 0.5-3 3-6 6-18 18-28 28-54 54-62 62-70+	10YR 6/3 10YR 7/3 8.25YR 6/4 7.5YR 5/6 7.5YR 4/6 7.5YR 6/4 10YR 6/4 10YR 6/4	m 3vcpl/2mcpr 2csbk/1vcpl 1&2mgr 1&2gr lgr m m	n.o. 4np0 4np0 1np0,3pf,br 1np0,3pf,br 1np0 & co n.o. n.o.	n.o. o;d d 2sc 2sc 2sc 2sc 1rsc	21.63 0.31

Tab. 1: Continued

Soil Profile; Maximum Phase	Flow Age ¹ (m.y.)	Horizon ²	Depth cm	Color (dry matrix)	Structure ²	Clay Films ²	Carbonate ² Morphology	Profile Index ³	Thickness- Index ³		
MP-14; 2	0.46 ± 0.08	A	0-0.2	10YR 7/3	sg	n.o.	n.o.	59.69	0.42		
		Avk	0.2-6	10YR 7/3 & 7.5YR 6/6	3vepl/2cpr	4np0	d,ss				
		Btk	6-14	7.5YR 5/6	3mgr/fabk	co	msm				
		Btkb1	14-32	6.25YR 5/6	3f & mabk/fpr	4npf; co	d, csm				
		Btkb2	32-48	6.25YR 5/4	3f & msbk	1npf; co	d, fsm				
		Btkyb1	48-76	6.25YR 5/8	3m & csbk	1npf; co	fsm				
		Btkyb2	76-103	6.25YR 4/6	3m & csbk	1npf, w	csm				
		Bkb	116-126	10YR 6/4	1m & csbk	n.o.	d, flsm				
		Coxb	126-143	10YR 6/4	m/msbk	n.o.	d, flsm				
		Btkb	143-150+	6.25YR 3/4	m	br, co	sm				
		MP-11; 2	0.56 ± .08	A	0-1	10YR 6/4	sg	n.o.	n.o.		
				Avk/Bvk	1-10	10YR 7/4 & 7.5YR 6/4	3vepl/2cpr	4np0	d		
				Bky	10-20	10YR 6/4	m/lgr	n.o.	d		
By	20-29			10YR 6/4	m	n.o.	n.o.				
Avyb	29-38			10YR 6/4	gr/fpl	4np0	n.o.				
Btyb1	38-52			7.5YR 5/4	3mabk	3npf; co	n.o.				
Btyb2	52-64			7.5YR 4/4	3m & cabk	4npf; co	n.o.				
Btkyb	64-86			7.5YR 4/6	2m & cpr	4npf; co	2ss				
Btkb	86-96			6.25YR 5/6	3cgr	4npk; co	s				
2Btkyb	96-111+			6.25YR 5/6	3cgr	4npf; co	2sm & sf & ss				
MP-29 ² ; 3	0.70 ± 0.06			Ab	0-2	10YR 6/4	sg	n.o.	n.o.		
				Avb	2-4	10YR 7/2 & 10YR 6/3	mpl	4np0	n.o.		
				ABb	4-7	7.5YR 6/3	f & mgr	3npf & co	n.o.		
		Btkb1	7-13	7.5YR 5/4	mgr	4npf & co	d				
		Btkb2	13-19	7.5YR 5.5/4	f&mgr/fsbk	4npf & co	d				
		Btkb3	19-27	7.5YR 6/4	mabk	3npf & co	sf, sm				
		K	27-33	7.5YR 7/4 & 10YR 8/3	m&cabk	n.o.	s				
		Btkb1 ⁸	33-48	10YR 7/3.5	fpr	4npf & co	sm				
		Btkb2 ⁸	48-60	10YR 6.5/3.5	m&fsbk	4npf & co	s				
		Btkb3 ⁸	60-69	10YR 6/3 & 10YR 5/4	gr/sbk	4npf & co	sf & sm				
		Btkb4	69-80	7.5YR 6.5/4	cabk	2&3npf&co fsm					
		Btkb5	80-118	10YR 7/3	sbk	2npf & co	sm				
		2Bkb	118+	10YR 7/3 & 7.5YR 5/6	gr	n.o.	s				
MP-8 ⁷ ; 3	0.73 to 0.90	A	0-0.2	10YR 6/3	m	n.o.	d	27.03	0.39		
		Avk	0.2-5	10YR 7/3	3vepl/2mcpr	4np0	ss				
		ABtk	5-8	8.75YR 6/3	1msbk/m	4np0/lpf	d				
		Btk	8-25	8.75YR 6/3	1mgr	co	d & s				
		Btkb	25-62	7.5YR 5/4	3f & mabk	4npf, br, co	d & s				
		Bkb	62-70+	10YR 7/3	m	n.o.	d & s				

Tab. 1: continued

MP-12; 3	0.75 ± .11	Avk A/Bk Bk1 Bk2 Btkb 2Btkyb	0-1.5 1.5-5.5 5.5-17.5 17.5-22.5 22.5-40.5 40.5-70+	10YR 7/3 7.5YR 4/6 10YR 7/4 10YR 7/4 7.5YR 5/4 7.5YR 6/4	3vcpl/2mcpr 3csbk 1csbk m/1csbk m m	4np0 4npf n.o. n.o. 4npf & co 4npf & co s	d d, csf d, flsf d, c2sm m2sm s
MP-24; 3	0.85 ± 0.05	A Avk BAK	0-1 1-6 6-8	10YR 8/3 10YR 8/3 & 10YR 7/3 7.5YR 7/6	sg 3vcpl/2mcpr lmp1	n.o. 4np0 3npf & co	d d d, fss
		Bk 2Btkb 2Bkb1 2Bkb2	8-30 30-72 72-103 103-110+	10YR 7/3 8.75YR 7/4 8.75YR 8/2 10YR 7/3	m m/1fsbk m sg	n.o. 4npf & co n.o. n.o.	d, cl & 2sm d, s d, d d, s
MP-25 ¹⁰ ; 3	0.99 ± 0.07	A Avk ABvk Btk Btkb Btkyb Btkb	0-0.2 0.2-4 4-6 6-15 15-22 22-40 40-111+	10YR 7/3 10YR 7/3 7.5YR 6/6 7.5YR 5/4 to 5/6 6.25YR 5/6 7.5YR 6/6 & 7.5YR 8/4 10YR 7/4 & 5YR 5/8	sg 3vcpl/2mcpr lmp1/m 3f&mgr 3m & fgr 3m&fabk/mgr	n.o. 4np0 4npf, br, co 2npf, co 4npf; co 2npf, co 3npf, co	d d d, 2ss, 2sm d, 3sf, ss d, 1sf, sm d, 3sc d, 2sc

¹ Radiometric (K/Ar) ages from TURPIN et al. (1984); ± values are 2 sigma error ranges.

² Horizon, structure, clay films, and carbonate-morphology notation from SOIL SURVEY MANUAL (1951) and revised Chapter 4, SOIL CONSERVATION SERVICE SOIL SURVEY MANUAL (1981) (Directive 430-U, Issue 1, SCS, Washington, D.C.) or BIRKELAND (1984). Modifiers of master horizons: b - buried genetic horizon; k - accumulation of calcium carbonate; ox - oxidized; t - accumulation of silicate clay; y - accumulation of gypsum; v - vesicular pores. Structure: (Type) sg - single grain; pl - platy; m - massive; gr - granular; sbk - subangular blocky; abk - angular blocky; cpr - columnar. (Grade) 1 - weak; 2 - moderate; 3 - strong. (Size) f - fine; m - medium; c - coarse; vc - very coarse. Clay Films: (Frequency) v - very few; 1 - few; 2 - common; 3 - many; 4 - continuous. (Thickness) n - thin; mk - moderately thick; k - thick. (Morphology) pf - ped face; po - pore; co - colloidal stain; br - bridges. Carbonate Morphology: (Distribution) d - disseminated; s - segregated; sf - filaments or threads; ss - seams; sm - soft masses; sc - concretions. (Abundance of mottles, concretions) 1 - few; 2 - common; 3 - many.

³ After HARDEN (1982).

⁴ Estimated age on basis of (1) interpretation or paleomagnetic data by D. Champion, pers. comm., U.S. Geological Survey, Menlo Park, California; (2) surface morphometric data (WELLS et al. 1985); and (3) carbon-14 date from rock varnish on basalt (DORN 1984).

⁵ None observed.

⁶ Radiometric age of underlying basalt flow; soil MP-7 partly developed in younger tephra, cinder deposits and colluvium that bury basalt flow derived from cone located 200-300 m upwind of site location.

⁷ Range in possible flow ages given for several flows in this profile area because complex field relations inhibit precise location of various flows.

⁸ Very dark gray to black coatings of probable manganese oxide-rich material on clasts present in buried Btk subhorizons.

⁹ Soil described in eolian deposits on basalt flow that subsequently was buried by three layers of cinders and tephra (WELLS et al. 1985).

¹⁰ Rubble zone at 60 cm depth.

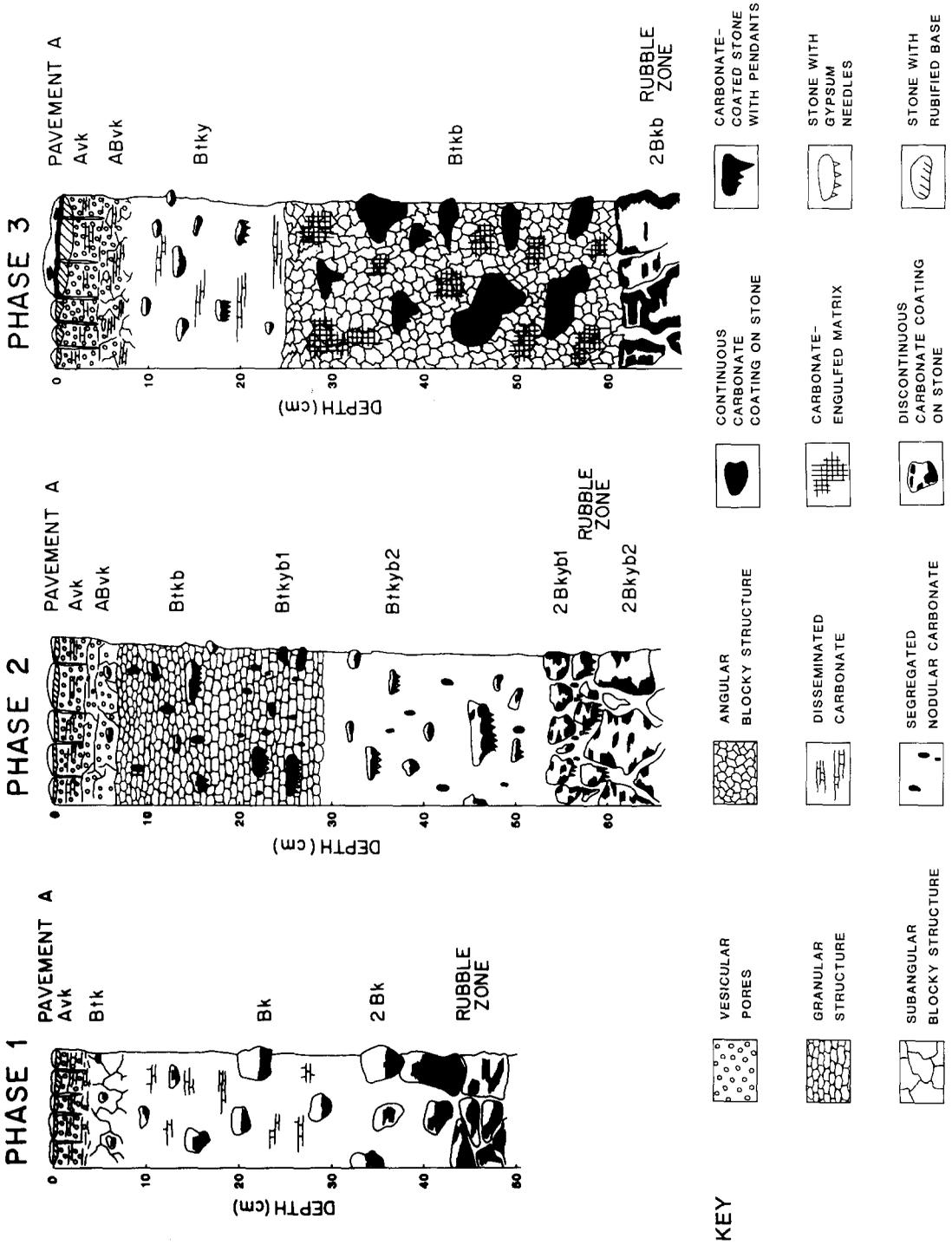


Fig. 2: Schematic diagrams showing the typical horizons and morphologies of phase 1 soils, phase 2 soils, and phase 3 soils. (See text for comments)

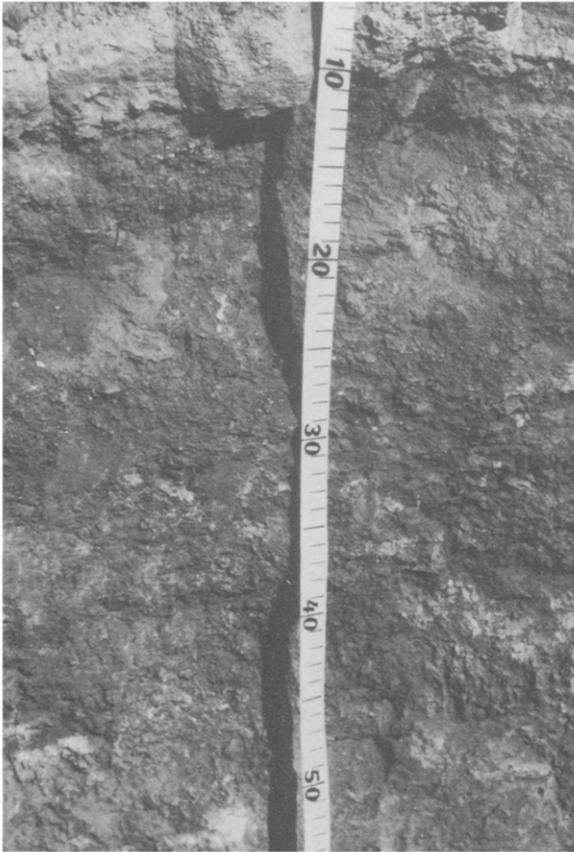
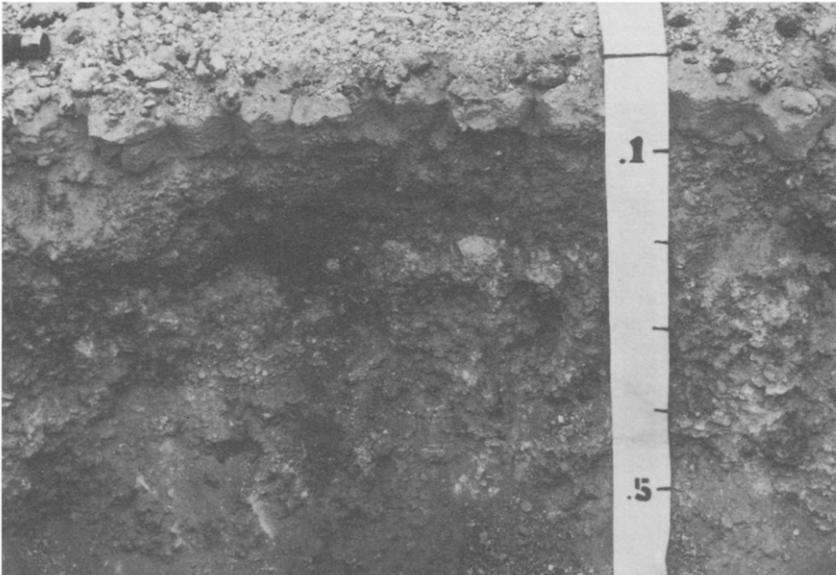


Photo 2: Soil profile (MP-27) developed in the loess deposits on a 0.27 ± 11 B.P.m.y. old flow. Note strong coarse platy to columnar structure of the pale vesicular A horizon and Stage I+ morphology of carbonate that has impregnated the argillic horizon. Scale in centimeters.

Photo 3: The soil profile (MP-25) developed in the loess deposits on a 0.9 ± 0.07 m.y. old flow. The top of the argillic horizon of the buried phase 3 soil occurs at a depth of 15 cm. Note massive impregnation of argillic horizon by carbonate. Scale in meters.



Tab. 2: LABORATORY DATA FOR SELECTED SOILS IN THE STUDY AREA; see Fig. 1 for location

Soil Profile	Horizon	Texture, <2 mm, %			pH (CaCl ₂)	EC (ms cm ⁻¹)	Fe ₂ O _{3d} (%)	Fe ₂ O _{3o} (%)
		Sand	Silt	Clay				
MP-5	A	42.2	56.9	0.9	7.8	0.25	0.75	0.06
	Avk	27.0	61.0	12.0	8.1	0.52	0.90	0.08
	Bk	39.5	57.4	3.1	8.3	0.56	0.86	0.05
	2Bk	53.3	43.6	3.1	7.9	0.45	0.64	0.05
MP-6	A	40.4	56.8	2.8	7.8	0.11	0.62	0.39
	Avk	29.3	59.1	11.6	8.4	0.41	0.99	0.07
	Btk	29.4	58.9	11.7	8.6	0.50	0.75	0.09
	Bk	46.2	49.6	4.2	8.4	1.27	0.69	0.09
MP-18	A	60.9	36.4	2.7	7.8	2.90	0.37	0.23
	Avk	35.3	48.7	16.0	7.9	2.09	1.00	0.54
	Bk1	67.3	23.4	9.3	7.9	8.14	0.80	0.53
	Bk2	68.3	27.7	4.0	7.8	8.59	0.64	0.49
	2Bky	68.4	25.4	5.7	7.8	4.53	0.46	0.48
MP-10	A	44.7	53.8	1.6	7.1	0.10	0.93	0.13
	Avk	20.5	60.2	19.3	7.8	0.21	1.24	0.12
	Btk	26.1	62.7	11.2	8.7	1.23	0.89	0.12
	Bk	24.0	61.7	14.3	8.8	2.68	1.16	0.14
	Bky1	30.9	59.6	9.7	8.5	7.23	0.87	0.28
	Bky2	27.2	61.7	11.04	8.6	5.53	0.85	0.23
MP-13	Avk	36.3	53.9	9.8	7.9	0.30	1.11	0.12
	Btk	43.4	49.5	9.1	8.1	0.38	0.75	0.10
	Bk1	54.2	41.7	4.1	8.0	7.27	0.75	0.13
	Bk2	49.9	48.0	2.1	7.9	6.04	0.41	0.06
MP-3	Avk	49.6	41.3	11.2	8.1	0.60	1.14	0.13
	Btk	36.4	40.8	22.8	8.4	4.09	1.28	0.21
	Bk	42.4	47.9	9.6	8.2	6.32	0.65	0.20
	Bky1	56.9	39.9	5.2	8.3	7.74	0.68	0.10
	Bky2	62.2	28.9	8.84	8.2	6.99	0.89	0.09
	Bky3	43.2	35.5	21.4	8.3	6.30	0.75	0.06
MP-14	A	39.9	51.3	8.8	7.8	5.09	0.28	0.18
	Avk	40.7	43.1	16.2	8.1	1.46	0.69	0.27
	Btk	26.2	50.7	23.0	8.1	1.95	0.34	0.18
	Btbk1	25.2	49.6	25.1	8.0	6.00	0.47	0.15
	Btkb2	30.6	39.9	29.5	8.1	5.49	0.61	0.34
	Btkyb1	36.7	48.1	15.2	8.0	7.25	–	–
	Btkyb2	33.1	41.0	25.9	8.1	6.22	0.61	0.36
	Btkb	42.3	42.6	15.1	8.1	4.12	0.70	0.40
	Bkb	44.7	43.4	11.9	8.2	3.65	0.56	0.28
	Coxb	39.3	50.8	9.9	8.1	–	0.55	0.27

Tab. 2: Continued

Soil Profile	Horizon	Texture, < 2 mm, %			pH (CaCl ₂)	EC (ms cm ⁻¹)	Fe ₂ O ₃ d (%)	Fe ₂ O ₃ o (%)
		Sand	Silt	Clay				
MP-11	A	54.1	39.7	6.2	7.5	0.10	0.86	0.22
	Avk/Bvk	30.3	55.6	14.1	7.8	0.19	1.16	0.20
	Bky	33.4	18.1	48.5	8.0	3.03	0.61	0.27
	By	65.6	20.9	13.4	7.7	2.96	0.47	0.30
	Avyb	55.5	23.4	21.1	7.6	3.70	0.63	0.16
	Btyb1	48.6	27.2	24.2	7.8	3.64	0.60	0.40
	Btyb2	26.6	34.1	39.4	7.6	8.70	0.81	0.19
	Btkyb	—	—	—	7.8	8.12	0.74	0.12
	Btkb	45.9	47.3	6.8	7.8	—	0.94	0.39
2Btkyb	27.0	29.8	43.2	7.8	7.75	1.21	0.27	
MP-1	A	77.1	20.5	2.4	7.5	0.27	0.65	0.09
	Avk	33.5	50.0	16.5	8.5	1.55	0.69	0.12
	ABk	43.4	41.0	15.6	8.3	1.84	0.87	0.15
	Btkyb1	56.6	32.4	11.0	8.2	3.63	0.61	0.16
	Btkyb2	60.3	29.8	9.9	8.0	2.48	0.63	0.15
	Btkyb3	32.0	44.2	23.8	8.1	3.77	0.75	0.16
	2Bkyb1	30.7	55.3	14.0	8.0	3.33	0.69	0.13
2Bkyb2	30.2	59.3	10.5	8.1	2.09	0.77	0.14	
MP-8	A	58.1	37.1	4.8	7.7	0.10	1.18	0.17
	Avk	45.2	45.8	9.0	7.9	0.10	1.33	0.17
	ABtvk	69.4	3.9	26.7	8.1	0.38	1.02	0.23
	Btk1	49.4	36.3	14.3	8.2	2.18	0.71	0.10
	Btk2	29.9	58.0	12.1	8.3	2.22	0.47	0.04
	Bk	70.4	19.3	10.4	8.8	1.67	0.39	0.06
MP-25	A	52.7	48.8	1.4	7.8	2.77	0.92	—
	Avk	42.5	44.0	13.5	7.9	2.81	1.11	0.19
	ABvk	48.7	40.9	10.4	8.1	2.37	1.06	0.27
	Btk	33.3	37.3	29.3	8.1	2.58	1.11	0.52
	Btkb	29.5	30.2	40.3	8.0	2.34	0.96	0.39
	Btkyb	39.0	38.3	22.7	7.8	2.57	0.53	0.05
	Btkb	28.0	40.8	31.3	7.8	2.67	0.70	0.13

The matrix of the B horizon, however, is often only weakly calcareous or noncalcareous. In many cases, coatings of secondary carbonate on the tops and sides of clasts are thinner and considerably less well indurated than the dense carbonate rinds commonly with thicknesses that exceed several mm that coat the undersides of clasts. The depth of secondary carbonate accumulation extends well below the top of the rubble zone. Secondary gypsum may also occur in the B horizon, usually as mantles of acicular euhedral crystals on carbonate coatings. Pockets of secondary gypsum may also be found in the soil matrix. Mottles and filaments of dark material presumed to be a type of secondary manganese oxide were observed in a few B subhorizons. A schematic diagram of the typical morphology of phase 2 soils is presented in Figure 2.

Values of the profile index for phase 2 soils range from 21.63 to 59.69 (Table 1). Part of the variability in the index value is related to profile thickness, as the range in the thickness-

normalized index is only 0.31 to 0.42 (Table 1). HARDEN & TAYLOR (1983) showed that soils exhibiting these values typically were developed in late to middle Pleistocene deposits. These data, therefore, indicate that the loess deposits in which phase 2 soils have formed are significantly older than the loess deposits in which phase 1 soils have formed, a conclusion consistent with the observation that phase 2 soils do not occur on flows less than ~ 0.18 M.y. old.

4.3 PHASE 3 SOILS

The soils on flows between 0.70 ± 0.06 M.y. and 0.99 ± 0.07 M.y. old, designated as phase 3 soils are strongly developed and usually possess Stage III calcic horizons and argillic B horizons that are engulfed by secondary carbonate (Table 1, 2; Photo 3). As with phase 2 soils, phase 3 soils may be buried by eolian deposits in which phase 1 soils have formed. In the phase 3 soil profiles, secondary carbonate is segregated as material that has apparently displaced clay plasma, as nodules, or as thick pendants on the bottoms of basaltic clasts. This more advanced carbonate morphology indicates that phase 3 soils are much older than phase 2 soils, a conclusion consistent with the ages of the flows beneath phase 3 soils. However, the large amount of clay in the argillic horizon, locally red colors, and abundant clay films show that the argillic horizon was very well developed prior to carbonate engulfment (Tables 1 and 2). A schematic diagram of the typical morphology of phase 3 soils is presented in Figure 2.

Values of the profile index of phase 3 soils range from 27.03 to 75.19 (Table 1). As with values noted for phase 2 soils, these values indicate a late to middle Pleistocene age for the loess parent materials of phase 3 soils. The higher of the two values also supports the interpretation that loess in which phase 3 soils occur is older than loess in which phase 2 soils have formed and is also consistent with the observation that phase 3 soils do not occur on flow surfaces younger than 0.70 M.y. old. The lower index value for a phase 3 soil is partly due to inclusion of a profile that is only 70 cm thick. This profile, however, exhibits massive impregnation of the Bt horizon by carbonate, which has destroyed or masked clay films, soil structure, and reddened soil color which has also lowered the value of the profile index. Moreover, the soil is located on a 3° to 4° slope, and therefore erosion of the argillic horizon prior to the most recent episode of loess deposition may have resulted in a decrease of the profile index value.

4.4 PHASE 4 SOILS

The soils present on Pliocene flows, designated as phase 4 soils, completely lack argillic B horizons but commonly exhibit large, partly buried blocky fragments of carbonate that are inferred to be the remnants of Stage IV-to-V petrocalcic horizons. These angular blocks occasionally exceed several decimeters in thickness. The carbonate is massive to strongly laminated, occasionally possesses interlaminated amorphous silica, and encloses scattered fragments of basalt. Parts of Pliocene flow surfaces are covered by weakly developed pavements of secondary carbonate clasts and detached carbonate coatings. Thin mantles of eolian material with weak soil development typically underlie these pavements.

5. CHEMISTRY AND MINERALOGY OF SOILS ON QUATERNARY FLOWS

5.1. CARBONATES AND SOLUBLE SALTS

Generally high electroconductivity (EC) and soil pH values reflect the presence of carbonates and soluble salts in soils on the Cima lava flows (Table 2). The only soils exhibiting EC values consistently lower than 0.50 mS cm^{-1} are those situated on the youngest lava flow (soil profiles MP-5 and MP-6, Table 2). Other phase 1 and all phase 2 and 3 soils have EC values that typically exceed 2.0 mS cm^{-1} and occasionally exceed 8.0 mS cm^{-1} . We infer that all soils in the study area contain at least some salts more soluble than calcium carbonate because a calcite-water sample yields an EC value of only 0.076 mS cm^{-1} . EC values are not clearly related to pH; however, values of EC are usually lowest in the A and vesicular A horizons and increase significantly with depth (Table 2), and maximum EC values typically occur above the Bk horizon or rubble zone, indicating that the locus of soluble salt accumulation is at a depth shallower than one meter.

Abrupt changes in EC at the Avk-Bk boundary or at the position of a buried soil suggest that major changes in soil texture with depth influence the position of some buried phase 2 soils (Table 1, MP-14, MP-27, MP-1) also indicates textural control of soluble-salt accumulation. However, in some cases, secondary gypsum was not detected at such boundaries, showing that other factors affect position of soluble-salt accumulation. The lack of accumulated gypsum in some soils (Table 1, MP-5, MP-6, MP-13) is attributed to the lack of a thick, clay-rich B horizon, although gypsum accumulation may occur within the rubble zone, below the maximum pit depth. The morphologic evidence and laboratory data show no trend of gypsum or soluble-salt accumulation with increasing degree of soil development (Tables 1 and 2). The data suggest instead that most gypsum and soluble-salt accumulation in soils of the study area probably occurred during the time when phase 1 soils were formed.

5.2. SECONDARY IRON OXIDES

Secondary iron oxyhydroxides have accumulated in the A and B horizons of soils on the Cima lava flows (Tables 2, 3). Increases in $\text{Fe}_2\text{O}_3\text{d}$ content range from 0.3 to 0.7% in the A horizon and between 0.1 and 1.1% in the B horizon. Soil $\text{Fe}_2\text{O}_3\text{o}$ content is typically much lower than the $\text{Fe}_2\text{O}_3\text{d}$ content although in a few soils, for unknown reasons, parent material $\text{Fe}_2\text{O}_3\text{o}$ content is atypically high. However, with the exception of only a few soils (MP-3, 8, 27), changes in $\text{Fe}_2\text{O}_3\text{o}$ with depth are not indicative of accumulation of pedogenic $\text{Fe}_2\text{O}_3\text{o}$. Because ferrihydrite ($\text{Fe}_2\text{O}_3\text{o}$) is the initially precipitated authigenic Fe phase prior to hematite formation (SCHWERTMANN & TAYLOR 1977), data for the Cima soils indicate that either: 1) ferrihydrite is transformed rapidly to hematite (as opposed to goethite) due to high summer soil temperatures, lack of soil moisture and alkaline soil pH, hence ferrihydrite content remains low, or 2) secondary Fe oxyhydroxides from external sources such as eolian dust are a significant component of accumulated iron oxyhydroxides (MCFADDEN 1982).

To evaluate the primary mechanisms of pedogenic Fe oxyhydroxide accumulation in these soils, the content of ferrous iron oxide (FeO_t), total ferric iron oxide (Fe_2O_3), and total Fe oxide ($\text{Fe}_2\text{O}_3\text{t}$) contents were determined for a phase 1 and phase 2 soil (MP-5 and MP-11, Table 3 and Fig. 3). In both soils, the FeO_t content and the $\text{FeO}_t/\text{Fe}_2\text{O}_3$ ratio is highest in the

Tab. 3: FERROUS IRON OXIDES (FeOt), FERRIC IRON OXIDES (Fe₂O₃) AND TOTAL IRON OXIDES (Fe₂O_{3t}) DATA FOR SELECTED PHASE 1 AND PHASE 2 SOILS

Soil Profile; Stage	Horizon	FeOt (%)	Fe ₂ O ₃	Fe ₂ O _{3t} (%)	FeOt/Fe ₂ O ₃
MP-5 (Stage 1)	A	1.52	3.22	4.91	0.47
	Avk	1.03	4.22	5.36	0.24
	Bk	1.09	3.89	5.10	0.28
	2Bk	1.10	3.18	4.40	0.35
MP-11 (Stage 1, A to By; Stage 2, Avyb to Btkb)	A	1.47	3.72	5.35	0.40
	Avk/Bvk	1.10	5.18	6.40	0.21
	Bky	0.65	3.38	4.10	0.19
	By	0.54	3.36	3.96	0.16
	Avyb	0.42	3.74	4.21	0.11
	Btyb1	0.55	4.42	5.03	0.12
	Btyb2	0.01	5.60	5.60	0.002
	Btkyb	0.20	6.38	6.60	0.03
	Btkb	0.34	6.42	6.80	0.05

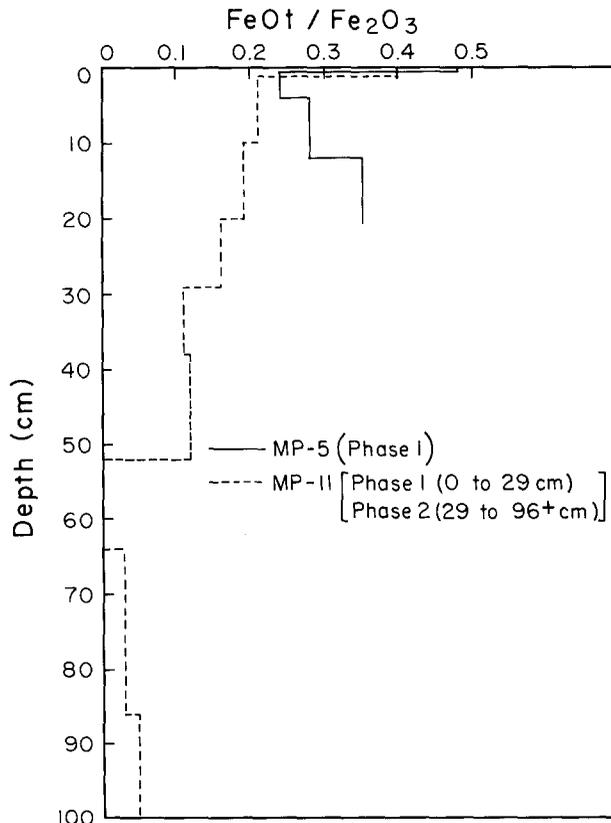


Fig. 3: The FeOt/Fe₂O₃ ratio with depth for a phase 1 and phase 2 soil. The boundary between the phase 1 soil and the buried A horizon of the phase 2 soil of profile MP-11 occurs at a depth of 29 cm and the top of the reddened B horizon occurs at a depth of 38 cm.

upper horizons and decreases slightly with depth. Net gains in Fe_2O_3 are attributed to subsequent additions of iron to the loessal parent materials from an external source, which is probably iron-bearing minerals in eolian dust. Indeed, PEWE et al. (1981) reported that eolian dust collected in central Arizona contains as much as 5.15% Fe_2O_3 . The buried argillic B horizon of MP-11, in contrast to overlying horizons, is characterized by a very low FeOt content, a very low FeOt/ Fe_2O_3 ratio, and a relatively high Fe_2O_3 content. These data indicate that a significant quantity of ferrous iron has been altered to ferric iron oxyhydroxides in the buried argillic horizon of the phase 2 soil; whereas, minimal iron alteration is indicated in the soil horizons of phase 1 soils.

Much of the increase in Fe_2O_3 and Fe_2O_3 contents may be due to additions of Fe oxides in dust. At least some authigenic ferric iron oxide accumulation in vesicular A and transitional AB or B horizons is suggested by the slight reddening. However, development of thin, discontinuous and reddened B horizons and accumulation of secondary Fe oxides on

Tab. 4: RELATIVE ABUNDANCES OF CLAY MINERALS IN SELECTED PHASE 1 AND PHASE 2 SOILS¹

Soil Profile	Horizon	illite/ mica	smectite	kaolinite	vermiculite	mixed layer ²	quartz	feld- spars
MP-5	A	2	1	1	1?	0	4	1
	Avk	3	1	2	0	0	4	1
	Bk	3	1	2	0	0	4	1
	2Bk	2	1	2	0	0	4	3
MP-6	A	2	1	1	1?	0	4	1
	Avk	3	1	2	0	0	2	4
	Btk	3	1	2	1?	0	3	1?
	Bk	3	2	2	0	0	3	1?
MP-10	A	3	3	1	0	0	3	1
	Avk	2	3	3	0	0	3	1
	Btk ³	3	3	3	1?	0	4	1
	Bk1	3	3	3	1?	0	3	1
	Bky1	2	3	3	1	0	4	1
	Bky2	2	3	3	1	0	4	1
MP-1	A	1	1	0	0	0	0	0
	Avk	2	1	2	0	0	4	2
	ABk	2	1	1	0	3	3	1
	Btyb1	2	1	2	0	4	2	1
	Btkyb2		1	2	0	3	2	1
	Btkyb3	3	1	3	0	2	3	1
	2Bkyb1	3	2	3	0	1	3	1
	2Bkyb2	3	2	2	0	1?	3	1

¹ 5 = predominant (> 90%); 4 = abundant (~ 50-90%); 3 = moderate (~ 10-50%); 2 = minor (~ 1-10%); 1 = trace; 0 = absent or undetected.

² Chief mixed-layer phase present is probably illite/hydroxy interlayered smectite, determined on basis of peak behavior during glycolization (increase from $7.85^\circ 2\theta$ to $7.75^\circ 2\theta$), K^+ saturation (no observed peak decrease) and 500°C heat treatment (collapse to $8.^\circ\theta$).

³ Trace chlorite also detected.

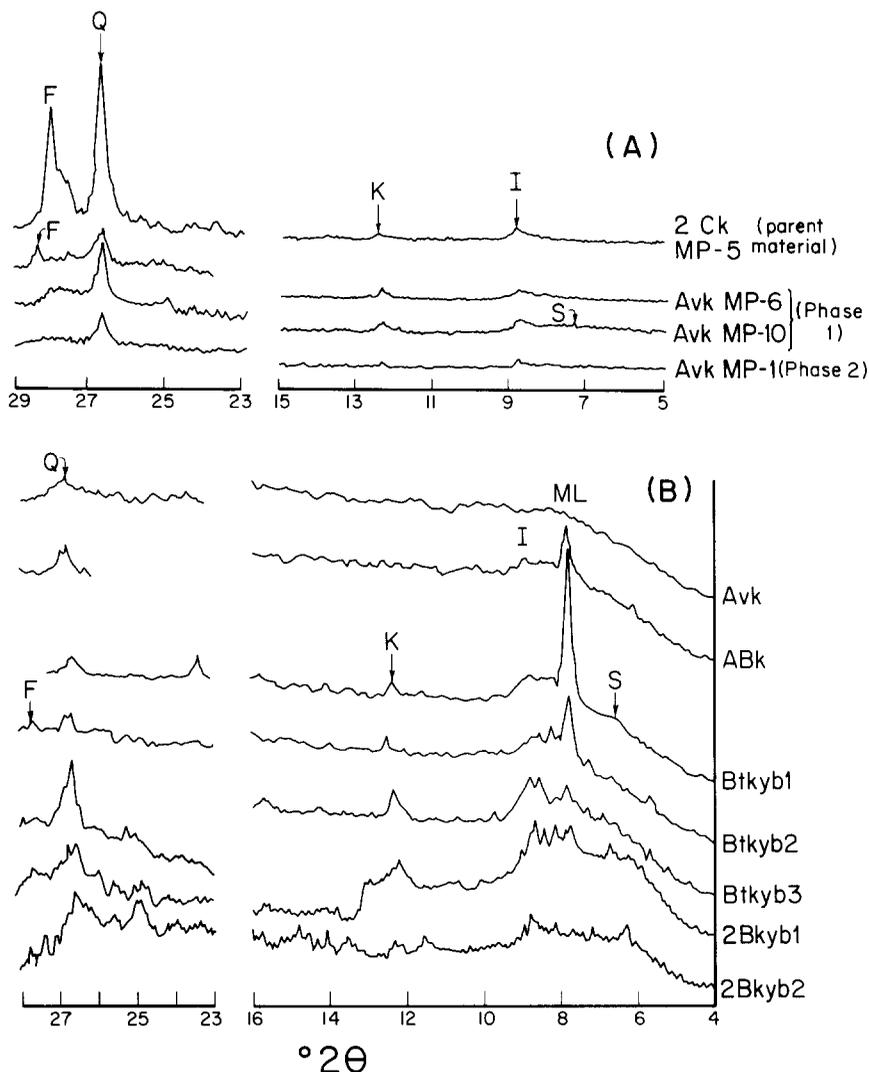


Fig. 4: X-ray diffractograms (Mg-saturated) of the vesicular A horizons ($< 2 \mu\text{m}$ fraction) of selected phase 1 (MP-6 and MP-10) and phase 2 soils (MP-1) and the loess parent materials ($< 2 \text{mm}$ fraction) (Ck). Note similarity of clay mineralogies of all vesicular A horizons of phase 1 soils and the vesicular A horizon associated with a phase 2 soil. X-ray diffractograms (Mg-saturated) of the subhorizons of a phase 2 soil (MP-1); a comparison with Fig. 4a shows the greater degree of phyllosilicate clay accumulation in phase 2 soils compared to phase 2 soils.

stone bottoms that project into the vesicular A horizon have been attributed to formation of small amounts of secondary ferric iron oxyhydroxides in the moist, weakly calcareous to noncalcareous microenvironment that is present below relatively large stones at the surface of desert soils (McFADDEN 1982).

5.3. CLAY MINERALS

X-ray diffraction studies of selected soils in the study area indicate that quartz, illite, and kaolinite are the most abundant minerals in the clay-size fraction of soils on the Cima flows (Table 4, Figs. 4a, b). Very minor amounts of smectite and a mixed-layer clay mineral may have accumulated in the thin Bt horizons of the weakly developed, phase 1 soils. In the argillic horizon of the strongly developed phase 2 soil, however, a mixed-layer clay mineral, kaolinite and illite become much more abundant relative to quartz (Fig. 4b). The behavior of this mixed-layer clay mineral during Mg, K, glycolization, and heat treatments suggests that it may be a hydroxy-interlayered smectite/illite.

As with the Fe-oxide analysis, the results of mineralogical analysis suggest that little chemical alteration has occurred in weakly developed phase 1 soils and that much of the increase in clay, quartz, and Fe oxides can be attributed to the presence of these minerals in accumulated eolian dust. These conclusions are consistent with the presence of abundant secondary carbonate and gypsum in phase 1 soils, which also must be attributed to derivation from external sources, given the lack of carbonate and sulfates in the underlying basalt flows and the lack of significant chemical alteration of basaltic rubble. In contrast to phase 1 soils, Fe oxyhydroxide and clay mineralogical data for the argillic horizon of the more strongly developed phase 2 soils do indicate a much greater degree of chemical alteration of soil parent materials, a conclusion consistent with field and petrographic evidence for more strongly altered basaltic rubble in phase 2 soils (MCFADDEN, unpublished data).

6. DISCUSSION

6.1. EVOLUTION OF PHASE 1 SOILS: VESICULAR A HORIZON AND B HORIZON DEVELOPMENT

An important aspect of the initial development of phase 1 soils is the development of the vesicular A horizon. Formation of this horizon is the result of entrapment and incorporation of eolian dust below the stone pavement, a process that increases the clay, silt, Fe oxides, and carbonate content of the coarser loess parent materials. We believe that the accumulation of clay in the loess profoundly affects the physical properties of the loess, primarily by increasing its shrink-swell capacity. Field observations show that the vesicular A horizon is characterized by a relatively uniform thickness and a pronounced nonorthogonal pattern of cracks (Photo 4). This pattern closely resembles the pattern of desiccation cracks formed in response to shrinkage of cohesive, fine-grained materials (COOKE & WARREN 1973). LACHENBRUCH (1962) attributed development of such nonorthogonal crack patterns (in contrast to orthogonal patterns) to simultaneous formation of cracks branching at obtuse angles in uniformly dried, homogenous, nonplastic materials. The textural and mineralogic character of the vesicular A horizon and its uniform thickness apparently provide the appropriate properties for development of nonorthogonal pattern of desiccation cracks.

Formation of the vesicular A horizon affects infiltration characteristics of the evolving phase 1 soil which, in turn, influences the nature of subsequent soil development. We artificially wetted vesicular horizons and noted that much infiltration occurred initially through inter-ped cracks, a process that probably accounts for the noncalcareous to weakly calcareous nature of the margins of vesicular A horizon peds. In addition, vesicular A horizon peds are markedly domed or convex-up (columnar structure) (Photo 4), a feature that can also be

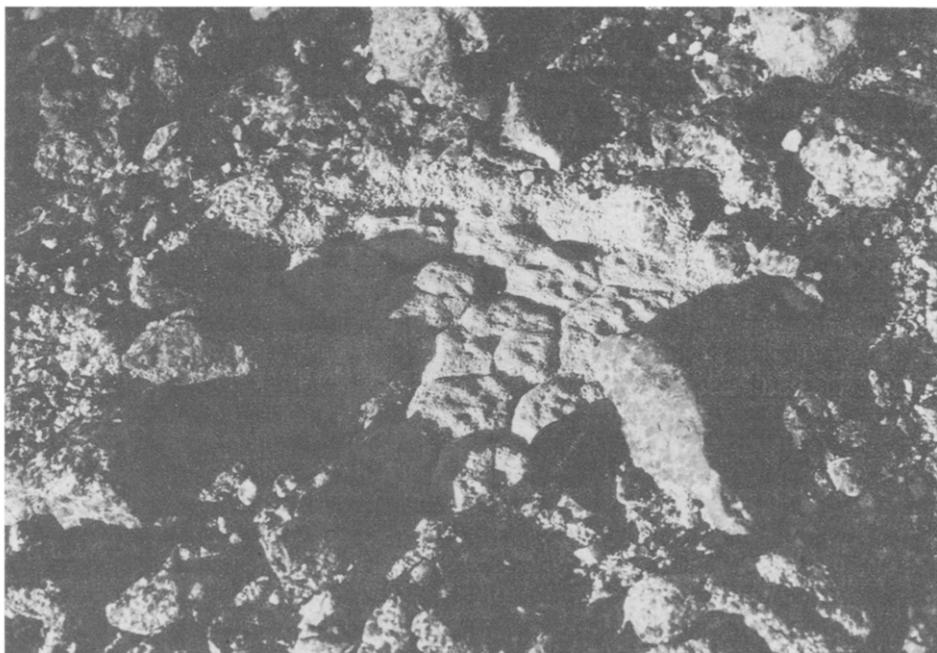


Photo 4: The nonorthogonal crack system of the vesicular A horizon, exposed after removal of closely packed stones of the pavement. Note convex, domed microtopography of blocks or peds (columnar structure). Lens cap diameter = 6 cm.

attributed to the development of desiccation-crack microtopography (COOKE & WARREN 1973). Our field observations also indicate this doming is seasonal in nature and occurs during dry soil conditions; increased soil moisture in the winter season apparently results in collapse of the domes. Doming of the peds may facilitate movement of water, clay, fine silt, and soluble materials to cracks during precipitation and wetting of initially dry soils. Continual accumulation of clay, silt, and Fe oxides, and some carbonate in the vesicular A horizon may enhance the magnitude of doming. Thus, the development of the clay-rich, domed vesicular A horizon may actually increase the overall rate of leaching of soluble salts and carbonate below the vesicular a horizon to subjacent B horizons.

We propose that the formation of the B horizon of phase 1 soils is related to development of the vesicular A horizon. Continuous deposition of eolian materials at a very low rate favors slow, upward growth of the vesicular A horizon. The most recently deposited eolian materials are present in the superjacent, thin A horizon. The horizon is slightly depleted in clay and carbonate, presumably due to rapid eluviation of these materials to the subjacent vesicular A or lower horizons. We conclude that accumulation of plasma in the vesicular A horizon eventually destroys previously formed vesicles in the base of the horizon, and a slightly reddened interior of the vesicular A horizon peds forms as slowly infiltrating moisture promotes chemical alteration of ferrous iron and accumulation of authigenic, bright red ferric iron oxyhydroxides. With continued growth of the cumulic A horizon, the interiors of peds eventually coalesce, forming a continuous and calcareous, reddened B horizon.

6.2. FORMATION OF STRONGLY DEVELOPED ARGILLIC HORIZONS OF PHASE 2 AND 3 SOILS

The process of cambic B horizon development described above could conceivably, over a long period of time, result in the formation of thick, cumelic B horizons. However, the thick argillic horizons on older Pleistocene flows formed at least partly in a pedologic environment different from the environment in which the cambic B horizons on late Pleistocene flows have formed. The thin Bt or transitional AB horizons are calcareous and lack abundant clay films. In contrast, the matrix of the argillic B horizon is largely noncalcareous; secondary carbonate has accumulated as soft nodules that clearly displace previously accumulated, silicate-clay plasma. This evidence indicates that the argillic B horizon formed over a long period of time in a noncalcareous environment that favored illuviation of silicate clay, presumably due to greater-than-present depths of leaching. Thus, the overlying, calcareous vesicular A horizon has formed much more recently and is not the eluvial horizon associated with development of the argillic B horizon.

The most recent major change in climate in the Mojave Desert occurred during the latest Pleistocene and early Holocene (SPAULDING et al. 1983). Soil-geomorphic studies and theoretical modeling of soil development in arid climates (McFADDEN & TINSLEY 1985, SCHLESINGER 1985) indicate that the Pleistocene-to-Holocene climatic change should have resulted in decreased depths of leaching and enhanced deposition of carbonates at shallower depths. Observed trends of soil development in the Cima volcanic field accord well with these findings. For example, at one site a well-developed, phase 2 soil (MP-29, Table 1) is buried by several late Pleistocene (~ 0.1 to 0.2 M.y.) tephra layers (WELLS et al. 1985). A significant quantity of secondary carbonate has accumulated in these tephra layers; however, the upper 7 cm of the buried phase 2 soil are noncalcareous and the soil is only weakly calcareous to a depth of 19 cm. These data indicate that, prior to burial by tephra, the soil had been forming in a climatic regime that was effectively wetter than the present climatic regime, because carbonate presently accumulates in the vesicular A horizon as well as in lower horizons. In addition, because the depth of soluble salt and gypsum accumulation is known to closely correspond with the maximum depth of leaching (DAN et al. 1982), the accumulation of soluble salts in the argillic horizon and gypsum in the Bk horizons of soils of the Cima volcanic field also indicates change to a shallower depth of leaching since formation of the argillic horizon.

Factors other than climatically induced changes in the depth of leaching might partly explain the apparent polygenetic character of the Cima soils. For example, the lower permeability and greater water-holding capacity of silty loess compared to sandy gravels on desert piedmonts should favor shallower leaching and carbonate/soluble salt accumulation in loessal parent materials (BIRKELAND 1984). Furthermore, the development of saline playas throughout the Mojave Desert during the Holocene has probably increased the carbonate and sulfate component of eolian dust as well as increasing the dust influx. PETERSON (1980), for example, determined that the presence of abundant soluble salts and rapid B-horizon development in soils on alluvial fans of the Panamint Mountains, 60 km north of the study area, was due to wind-blown salts and clays derived from the adjacent Death Valley playa. WELLS et al. (in press) also noted the presence of gypsiferous soils in Holocene gravelly piedmont deposits of the Soda Mountains, 25 km northwest of the Cima volcanic field. These results show that the leaching regime in the hot, arid climate of the Mojave Desert is sufficiently shallow and the rate of sulfate influx sufficiently high to cause accumulation of gypsum and soluble salts in relatively permeable gravelly soils. Therefore, accumu-

lation of abundant gypsum and carbonate at shallow depths in the soils of the study area are attributed to both a shift to shallower depths of leaching and a probable increase in the supply of sulfate and carbonates in dust during the latest Pleistocene to early Holocene.

The Fe oxide content and clay mineralogy of the Cima soils also argue for polygenetic soil development. As discussed previously, the $\text{FeOt}/\text{Fe}_2\text{O}_3$ ratios strongly suggest depletion of primary ferrous oxides and associated accumulation of secondary ferric oxides in the maximally developed part of the argillic horizon. In addition, a mixed-layer clay mineral interpreted to be authigenic has also accumulated in the argillic horizons. These pedogenic features, as well as presence of translocated clay-rich, carbonate-poor argillic horizons, are consistent with a greater magnitude of in-situ chemical weathering favored by more intense leaching during previously effectively moister climatic regimes.

6.3. DEVELOPMENT OF STONE PAVEMENTS

Stone pavements constitute the surface layer of all loess deposits on lava flow surfaces. The Cima pavements are composed of relatively well sorted clasts that are one particle thick, and coated by a dark layer of varnish. The Cima pavements bear a strong resemblance to stone pavements on piedmonts of other arid regions (COOKE & WARREN 1973). The origin of stone pavements has been attributed to several processes which include deflation of fines and concomitant accumulation of coarser lag, and shrink-swell-induced upward movement of stone through a clay-rich B horizon (COOKE & WARREN 1973). However, pavements present on the youngest loess deposits that bury flow surfaces have developed in association with soils that lack argillic horizons; thus, the formation of these pavements cannot be attributed to upward movement of stones through an argillic horizon. Furthermore, the presence of well-developed stone pavements associated with young, weakly developed soils formed in deposits that bury much older, phase 2 and 3 soils shows that the stones of the pavement can be maintained continually at the surface. This conclusion is supported by the fact that, on flat surfaces of middle Pleistocene and older age, there is not now nor could there have been in the last 0.18 M.y. exposed bedrock that would be the only possible lateral source of stones for the existing pavement. Therefore, the existing pavements have been derived from pavements formerly associated with the well-developed, but presently buried, soils. Apparently, the rate of loess deposition on such flow surfaces is slow enough to permit gradual incorporation of eolian dust and concomitant vesicular A horizon development beneath existing pavements; consequently, the stone pavement is not buried but instead is lifted and separated vertically from the surface of the well-developed argillic horizon during slow, progressive loess deposition.

6.4. SOIL DEVELOPMENT AND EVOLUTION OF LANDSCAPES ON FLOW-SURFACES

The most significant differences among soil phases can be attributed to time-dependent changes in soil development and to variations in climatic regimes during the Quaternary. Factors other than time or climatic change locally may influence the character of soil development. For example, variations in topography on flow surfaces affect the rate of eolian deposition and the initial permeability of the underlying rubble zone (WELLS et al. 1985), and such geologic controls are probably responsible for variations in soil development within

a single soil phase. In no case, however, do such variations exceed or even approach those variations in soil development that distinguish soil phases in the Cima volcanic field.

The existence of only four phases of soil development is unusual considering the broad range of flow ages. This four-phase sequence is strongly influenced by rates and timing of eolian deposition. For example, the presence of similarly developed soils on flows as young as 16,000 yrs or as old as 0.14 ± 0.04 m.y. indicates that the parent materials of the soils are of essentially the same age. Thus, deposition of the most recent loess mantle in which phase 1 soils have formed post-dates 16,000 yrs B.P. and represents the only major loess event of sufficient magnitude during at least the past 0.14 M.y. to bury large parts of late Pleistocene flow surfaces. This postulated loess event, as well as previous loess events, may be due to episodic exposure of playa floors and distal piedmont areas during interpluvials of the Quaternary (WELLS et al. in press). The high surface roughness of unmodified to weakly modified flows strongly favors entrapment of loess.

The moderately developed, buried phase 2 soils of older flows represents a long period of soil development on an older loess mantle. This indicates a long period of surface stability occurred prior to the most recent eolian event. However, strong argillic horizon development may be due partly to more intense leaching in previous Pleistocene climates. Flow surfaces in the Cima volcanic field that have phase 2 soils are geomorphically stable exhibiting the most extensive stone pavements. In contrast to these flow surfaces, flow surfaces with phase 3 soils are more dissected than younger flow surfaces and are eroded on flow margins exposing carbonate-coated rubble and bedrock. This dissection requires about 0.4 to 0.7 million years to develop. Impregnation of the argillic horizons with pedogenic carbonate in strongly developed phase 3 soils inhibits infiltration and induces significant surface runoff (WELLS et al. 1985). Attainment of phase 3 soil development is the inevitable result of continued carbonate and clay accumulation favored by prolonged soil development in semiarid and arid environments with available sources of eolian calcareous dust. The increasing magnitude of dissection prohibits continued growth and development of increasingly thicker, well-developed argillic horizons.

Once a runoff-dominated hydrologic regime is attained, the deposition of thick eolian mantles during episodes of loess accumulation is inhibited. However, at one site, a one-meter thick phase 1 soil was encountered on such a surface. Deposition of loess in channels established on the surface prior to loess deposition may account for occasionally thick loess accumulation. In most cases, however, 15-30 cm thick loess mantles with phase 1 soils typify these surfaces, in contrast to the much thicker loess mantles with phase 1 soils that always occur on late Pleistocene flow surfaces (Table 1). Because the deposition of relatively thin mantles would be expected on any flow surface that has attained the surface morphology associated with phase 3 soils, it is possible that thin mantles may have been deposited on the surfaces of > 0.7 M.y. old flows more than once during the past 0.5 to 0.3 M.y. Strong argillic B horizon development within such a thin mantle during a subsequent long period of stability could thicken the Bt horizon by superposing the buried B horizon and the slowly forming, overlying argillic B horizon. Superposed Bt horizons might well be difficult to recognize on the basis of field observations; however, some of the variation in particle size and content with depth in these horizons (for example, MP-25) might be attributable to superposition. Presumably, the development of the argillic B horizon in a thin mantle would accelerate surface erosion as discussed above.

Extensive basalt flows older than 0.99 ± 0.07 M.y. B.P. and younger than 3.3 M.y. B.P. have not been located in the Cima volcanic field; hence, we do not observe soils which might preserve the continuous Stage IV calcic horizons that would probably evolve on flows of this

age range. The presence of abundant fragments of massive, locally laminated carbonate on the surfaces of Pliocene flows indicates, however, that Stage IV to Stage VI calcic horizon morphology had been attained on these older flow surfaces. The eventual destruction of such a soil probably occurred after the argillic horizon, presumably impregnated by abundant carbonate, was completely stripped, exposing and eventually destroying the petrocalcic horizon. The weakly developed soils in the most recently deposited loess partly bury the rubbled petrocalcic horizon on Pliocene flow surfaces.

7. CONCLUSION

Soil-stratigraphic studies of the eolian mantle developed on flow surfaces of the Cima volcanic field indicate episodic loess deposition with intervening periods of soil formation. The development of reddened, clay and Fe oxyhydroxide-rich B horizons in loess deposits on flows of the Cima volcanic field occurred when rates of deposition decreased sufficiently over long periods of time to enable soils to develop. During other periods of time, rates of loess deposition increased sufficiently to cause partial burial of flow surfaces and soils formed in older loess deposits. The development of soils under conditions of relatively very slow eolian dust deposition results in formation of cumulic soils, a process also noted in the desert and desert-fringe soils in Israel (DAN 1965, YAALON & DAN 1974).

We attribute the episodic nature of loess deposition to climatic changes involving increasing aridity and disappearance of pluvial lakes. The development of playas and widespread exposure of loose unconsolidated fines in response to this climatic change was apparently the source for the eolian mantle developed on the basalt flows. Eolian depositional rates would be reduced under more humid climatic regimes which restrict sediment availability, enhance vegetational cover, and promote soil development. Deposition of loess and an increasing degree of soil development in loessal parent materials greatly influence the landscape evolution of volcanic flow surfaces in the Cima volcanic field, a conclusion that has significant implications for landscape evolution on other surfaces in regions subjected to episodic loess deposition. Results of this study also indicate that the well-developed phase 2 soils may have formed under climatic conditions different than those presently existing in the Mojave Desert. Thus, soils developed in eolian mantles on volcanic flows serve as valuable indicators of Quaternary paleoclimates in the Mojave Desert of the southwestern United States.

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