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CALDEP: A REGIONAL MODEL FOR SOIL $CaCO_3$ (CALICHE) DEPOSITION IN SOUTHWESTERN DESERTS

G. M. MARION,[1] W. H. SCHLESINGER,[2] AND P. J. FONTEYN[3]

Our objective was to develop and validate a regional model for $CaCO_3$ deposition in desert soils of the southwestern United States. There were five major components in the simulation model: a stochastic precipitation model, an evapotranspiration model, chemical thermodynamic relationships, soil parameterization, and a soil water and $CaCO_3$ flux model.

For the present climate, a cold-dry Pleistocene climate, and a cool-wet (summer) Pleistocene climate, the model predicted a shallower depth for the calcic horizon than was found in field soils. However, the model was compatible with field soils if one assumed that most pedogenic carbonate formed during a cool-wet (winter) Pleistocene climate. The model was highly sensitive to the frequency of extreme precipitation events and to soil water-holding capacity. The biotic factor played an important role in $CaCO_3$ deposition through its control of soil CO_2 concentrations and evapotranspiration rates. The range in predicted $CaCO_3$ deposition rates agreed with the rates for most field studies (1 to 5 g/m^2/yr); also, the model predicted an increasing rate of $CaCO_3$ deposition with increasing precipitation, which agreed with field studies. The model is a valuable research tool for evaluating the role of state factors on soil $CaCO_3$ deposition.

Calcium carbonate (caliche) soil horizons are common features of southwestern deserts. State factors that can influence the formation of soil calcic horizons include climate, parent material, time, topography, and biota. The depth of $CaCO_3$ horizons is strongly dependent on soil water flow and increases with increasing mean annual precipitation (Arkley 1963; Jenny 1980). Through its effect on evapotranspiration, temperature also plays an important role in controlling water flow in the soil (Arkley 1963; Ahmad 1978; McFadden 1982). Wind plays a critical role in the rate of formation of calcic horizons, because wind-borne dust and dissolved constituents in precipitation are considered the dominant sources of calcium for deposition as $CaCO_3$ in noncalcareous desert soils (Brown 1956; Reeves 1970; Gardner 1972; Gile et al. 1981; McFadden 1982; Schlesinger 1985). Soils formed on calcareous parent materials accumulate pedogenic $CaCO_3$ at higher rates than soils formed from noncalcareous parent materials (Lattman 1973; Schlesinger 1982, 1985). Parent material also controls, to a large extent, the water-holding capacity (WHC) of soils; the WHC, in turn, controls the depth of wetting and $CaCO_3$ deposition (Arkley 1963; Stuart and Dixon 1973; Ahmad 1978; Gile et al. 1981; McFadden 1982). The biotic factor can influence $CaCO_3$ deposition through its effects on (1) soil CO_2 concentrations, which largely control soil pH and $CaCO_3$ solubility, and (2) evapotranspiration (Arkley 1963; Ahmad 1978; Gile et al. 1981; McFadden 1982, Schlesinger et al. 1986).

A simulation model is the ideal way of integrating these diverse state factors into a holistic conceptual model of the formation of soil calcic horizons. Although several published simulation models consider $CaCO_3$ deposition and dissolution, most are short-term models not specifically designed for evaluating the long-term $CaCO_3$ deposition process (Dutt et al. 1972; Robbins et al. 1980; Dudley et al. 1981). The first model specially designed for predicting $CaCO_3$ deposition in soils was developed by Arkley (1963), who used precipitation, potential evapotranspiration, and soil WHC to evaluate the average amount of water passing through given horizons over an annual cycle. These water fluxes, coupled with internal calcium availability and solution $CaCO_3$ chemistry, were used to estimate $CaCO_3$ deposition. Using a similar water budget model, Ahmad (1978) developed a regression model relating caliche depth to soil moisture penetration and soil texture. A modification of

[1] Systems Ecology Research Group, San Diego State Univ., San Diego, Calif. 92182.

[2] Dept. of Botany, Duke Univ., Durham, N.C. 27706.

[3] Dept. of Biology, Southwest Texas State Univ., San Marcos, Tex. 78666.

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the Arkley model developed by McFadden (1982) included external calcium sources and a more sophisticated treatment of $CaCO_3$ chemistry to predict soil $CaCO_3$ deposition. The latter three models rely on the evaluation of several years' of climatic data to determine average annual leaching indices. Extreme events (e.g., storms), which may play a critical role in $CaCO_3$ transport, were not explicitly considered. The Arkley, Ahmad, and McFadden models implicitly assumed that calcium moved through the soil as mass flow with water. An alternative model evaluated diffusion as the dominant mechanism for $CaCO_3$ deposition (Marcoux 1978), but, based on present-day solution concentrations of calcium, bicarbonate, and pH, the diffusion model was invalid (Marcoux 1978).

Although correlations between present climate and the depth of the $CaCO_3$ horizons are strong (Arkley 1963; Jenny 1980), this does not imply a causal relationship. Many features of desert soil profiles, such as $CaCO_3$ and clay horizons, probably formed under earlier "wetter" climates (Gile et al. 1966; Nettleton et al. 1975; Bachman and Machette 1977; McFadden 1982). During the late Pleistocene (20 000 YR BP) in the Southwest, woodlands existed over much of the present-day desert; workers have implied a major climate change to explain these changes in vegetation (Wells 1963, 1979, 1983; Van Devender and Spaulding 1979; Galloway 1983).

The nature of the late Pleistocene climatic change in the Southwest is controversial. Galloway (1983) has hypothesized that the late Pleistocene climate in the Southwest was 10°C colder, with 80% of present precipitation (cold-dry hypothesis). In comparing the present distribution of woodlands and deserts, Wells (1966) hypothesized that late Pleistocene climates in the Southwest were cooler and wetter than the present climate (cool-wet hypothesis). For example, present woodlands average 45 cm of precipitation, while deserts average 27 cm (Wells 1966). Also, an 800-m elevational decrease in the woodland zone was postulated (Wells 1966); assuming an adiabatic lapse rate of 6°C/1000 m, this implied an approximately 5°C decrease in temperature. Wells (1979) has postulated that the increase in precipitation occurred in the summer months. To the contrary, Van Devender and Spaulding (1979) have postulated that the increased precipitation occurred in the winter.

The basic objectives of our study were to (1) develop a $CaCO_3$ deposition model (CALDEP) that was (a) an event-based process model, (b) easily parameterized for sites, and (c) useful for long-term simulations, (2) validate the model with field data, (3) consider the roles of state factors in controlling $CaCO_3$ deposition, and (4) use the model to examine the various hypothesized Pleistocene climates.

THE THEORETICAL MODEL

There were five major components to the model for $CaCO_3$ deposition in desert soils: a stochastic precipitation model, an evapotranspiration model, chemical thermodynamic relationships, soil parameterization, and water and $CaCO_3$ fluxes.

The stochastic precipitation model

Data from seven sites were used to quantify the stochastic precipitation model (Fig. 1, Table 1). These sites were chosen because they covered a broad range in both total annual precipitation and seasonal distribution of precipitation. In general, total annual precipitation and the proportion of summer precipitation increased from west to east (Table 1). Using the mean precipitation for each month, we separated the climate of each site into two or three seasons. In general, winter precipitation is derived from synoptic fronts from the Pacific Ocean, and summer precipitation is due to local scattered thunderstorms; exceptions include Yuma, which has a summer drought, and Phoenix and Tucson, which have a spring drought (Table 1).

Precipitation data were from NOAA (National Oceanic and Atmospheric Administration) publications: *Local Climatological Data,*

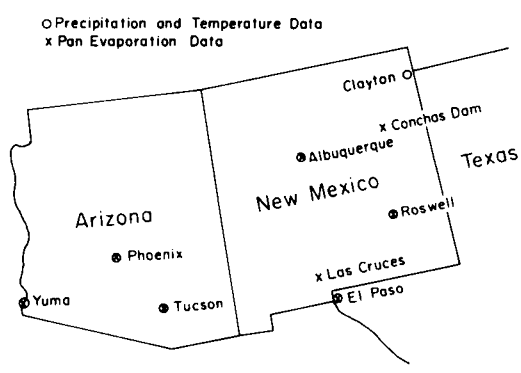

FIG. 1. The sources of climatic data for parameterizing the CALDEP model.

TABLE 1

A summary of the precipitation regimes for the selected sites

Site	Precipitation seasons	Annual precipitation, %	Annual precipitation, cm
Yuma, Arizona	August–March	87.7	8.5
	April–July	12.3	
Phoenix, Arizona	October–March	56.6	18.9
	April–June	7.8	
	July–September	35.6	
Tucson, Arizona	October–March	42.6	28.4
	April–June	7.1	
	July–September	50.3	
Albuquerque, New Mexico	November–June	46.0	21.1
	July–October	54.0	
El Paso, Texas	November–May	31.6	21.6
	June–October	68.4	
Roswell, New Mexico	November–April	25.3	31.6
	May–October	74.7	
Clayton, New Mexico	October–April	25.1	37.8
	May–September	74.9	

Monthly and Annual Summary. These records are available from the National Climatic Data Center in Asheville, North Carolina. Forty-eight months (January 1980 through December 1983) of precipitation data were used to develop the frequency distributions for interarrival time (i.e., the number of days between storms) and daily precipitation.

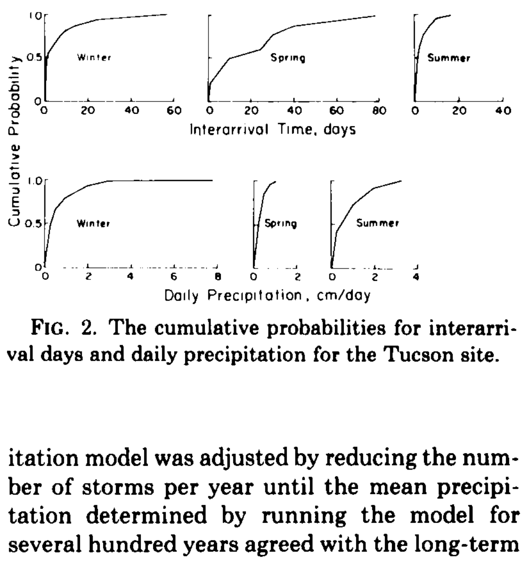

FIG. 2. The cumulative probabilities for interarrival days and daily precipitation for the Tucson site.

For Tucson, storm frequency, which is inversely proportional to interarrival time, decreased in the following order: summer > winter > spring (Fig. 2). Daily precipitation generally decreased in the order: summer > winter > spring (Fig. 2). Thus, in Tucson the three precipitation seasons were: summer with frequent, high-intensity storms; winter with less frequent and less intense storms; and spring with infrequent, low-intensity storms.

A random number generator was used in the model to select the interarrival time and daily precipitation based on cumulative probability distributions for each site and season (Fig. 2). A linear interpolation formula was used to estimate interarrival time and daily precipitation from tabular data. The maximum interarrival time and daily precipitation in the model were those found in the 4-yr data record. The consequences of terminating the cumulative probability distributions at these maxima and using only a 4-yr data base will be discussed later.

The 1980 to 1983 period that was used to develop the frequency distributions was generally wetter than normal. The stochastic precipitation model was adjusted by reducing the number of storms per year until the mean precipitation determined by running the model for several hundred years agreed with the long-term mean calculated by NOAA for each site. The simulation model accurately predicted the mean annual precipitation (see Tables 1 and 2). In general, the 40-yr weather-record mean was slightly lower than the simulation model prediction (Table 2), but this simply reflected the generally lower precipitation during the 1943 to 1982 period compared with the long-term record mean. The variances of the means, as judged by the standard deviations, were similar for the simulation model and the weather record (Table 2). The stochastic precipitation model accurately simulated both the mean and variance of annual precipitation.

TABLE 2

A comparison of the annual precipitation (mean ± 1 SD) predicted by the simulation model (100-yr run) with a 40-yr weather record (1943–1982)

Site	Simulation model, cm	Weather record, cm
Yuma, Arizona	8.0 ± 3.4	6.8 ± 3.6
Phoenix, Arizona	18.9 ± 6.5	17.7 ± 7.3
Tucson, Arizona	28.0 ± 9.9	28.1 ± 7.5
Albuquerque, New Mexico	20.8 ± 5.0	20.0 ± 5.1
El Paso, Texas	22.0 ± 6.4	20.1 ± 7.2
Roswell, New Mexico	30.3 ± 9.8	28.0 ± 4.0
Clayton, New Mexico	38.6 ± 8.8	37.6 ± 9.7

The evapotranspiration model

Actual evapotranspiration was calculated in three steps. First, potential evapotranspiration was calculated using Thornthwaite's equation (Thornthwaite 1948). Second, Thornwaite's potential evapotranspiration was converted to pan evaporation using a derived, empirical relationship. Third, actual evapotranspiration was calculated as a function of soil moisture and pan evaporation. For Pleistocene simulations at lower temperatures, a special adjustment was used to reduce pan evaporation before calculating actual evapotranspiration. If pan evaporation data were available for a given site, then steps 1 and 2 were eliminated.

Thornthwaite's equation for potential evapotranspiration was used, because it was a function only of mean monthly temperatures, which were readily available in the NOAA publications. The Thornthwaite potential evapotranspiration (TPET, $g/cm^2/m$) was calculated as follows

$$\mathrm{TPET} = 1.6(10\,T/I)^a \quad (1)$$

where T is mean monthly temperature (°C), I is an annual heat index, and a is a constant. The heat index (I) was calculated as follows

$$I = \Sigma_{i=1}^{12}\,(T_i/5)^{1.514} \quad (2)$$

where i is the month, and T_i is the mean monthly temperature (°C). The coefficient a was calculated from the empirical equation

$$a = C_1 I^3 + C_2 I^2 + C_3 I + C_4 \quad (3)$$

where $C_1 = 6.75 \times 10^{-7}$, $C_2 = -7.71 \times 10^{-5}$, $C_3 = 0.01792$, and $C_4 = 0.49239$. An empirical relationship was used to adjust TPET to pan evaporation (PEV). This relationship was developed from PEV data from seven sites including Yuma, Phoenix (Mesa), Tucson, Las Cruces, El Paso (Ysleta), Roswell (Bitter Lakes), and Conchas Dam (Fig. 1) (NOAA 1982a,b,c). The PEV/TPET ratio was strongly and negatively correlated with mean monthly temperature (Fig. 3); also, the data were sorted into separate relationships for the January to July period and the August to December period (Fig. 3). Tests of the regression coefficients indicated that all were significantly different from 0.0 at the 95% confidence level.

The relationship of actual evapotranspiration to potential evapotranspiration (pan evaporation) as a function of soil moisture is a subject of some controversy. Veihmeyer and Hendrickson (1955) argued that moisture was lost at the potential rate across the whole range of available water between field capacity (0.01 MPa) and permanent wilting point (1.5 MPa); Thornthwaite and Mather (1955) argued, however, that moisture loss was a linear function of moisture content across this range (Fig. 4). Actual evapotranspiration from a *Larrea tridentata* site at the Jornada Desert Long-Term Ecological Research (LTER) site near Las Cruces was estimated from soil thermocouple psychrometer data after converting soil moisture potential to moisture content with a moisture characteristic curve (Schlesinger et al. 1986). Pan evaporation was measured with a Class A pan. These data fell between the Veihmeyer-Hendrickson and

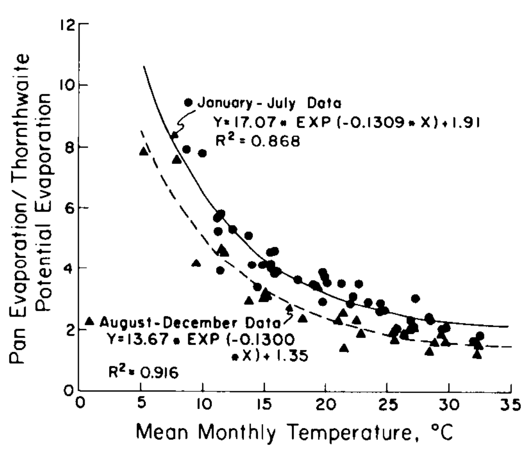

FIG. 3. The pan evaporation/Thornthwaite potential evapotranspiration ratio as a function of mean monthly temperature.

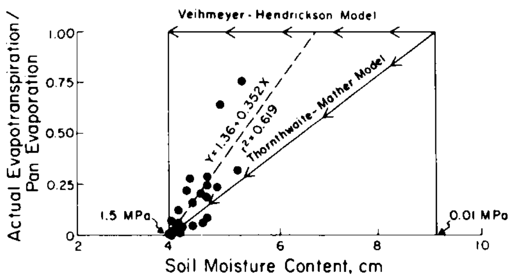

FIG. 4. The relationship of the actual evapotranspiration/pan evaporation ratio as a function of soil moisture content.

Thornthwaite-Mather models (Fig. 4); these results agreed with previous work, which has generally shown that actual evapotranspiration fell between the extremes of the Veihmeyer-Hendrickson and Thornthwaite-Mather models (Hanks and Ashcroft 1980). A linear regression for our data intersected the potential line (ratio = 1.0) dividing the available water range into two regions (Fig. 4). The simulation model assumed that in the upper 45% of the range, water was lost at the potential rate; in the lower 55% of the range, water loss was a linear function of soil moisture (Fig. 4). The total profile water content was used as the soil moisture base, rather than the water content of individual horizons, in evaluating moisture loss in the simulation model.

Simulations of Pleistocene climate were run assuming that the mean annual temperature was 5 or 10°C lower than at present. A model adjustment was necessary to account for the effect of this temperature decrease on PEV. Calculated annual PEV, using the previously discussed model, was strongly correlated with mean annual temperature (Fig. 5). Measured PEV could not be used directly because most of the sites have missing data. However, graphical estimates for the missing months plus the measured PEV data gave PEV estimates for the seven sites ranging from 220 to 290 cm, which agreed reasonably well with the calculated range of 230 to 300 cm. The average deviation between measured and calculated PEV was 20 cm. Monthly PEV was adjusted by multiplying by the ratio of the equation in Fig. 5 evaluated at the lower mean annual temperature and at the present mean annual temperature. For example, the calculated PEVs at 15 and 20°C were 234 and 261 cm/yr, respectively (Fig. 5); the mean monthly PEV was multiplied by 0.90 (234/261) to account for this hypothetical 5°C decrease in temperature. Because only the three warmest sites were used in the Pleistocene simulations, all temperature extrapolations with one exception fell within the range of the data (Fig. 5).

Chemical thermodynamic relationships

The chemical equilibrium equations used in this study include the following

$$\frac{(CO_2)}{P_{CO_2}} = K_1 \qquad (4)$$

$$pK_1 = 1.14 + 0.0131\,T \qquad (5)$$

$$\frac{(H^+)(HCO_3^-)}{(H_2O)(CO_2)} = K_2 \qquad (6)$$

$$pK_2 = 6.54 - 0.0071\,T \qquad (7)$$

$$\frac{(H^+)(CO_3^{2-})}{(HCO_3^-)} = K_3 \qquad (8)$$

$$pK_3 = 10.59 - 0.0102\,T \qquad (9)$$

$$(Ca^{2+})(CO_3^{2-}) = K_4 \qquad (10)$$

$$pK_4 = 7.96 + 0.0125\,T \qquad (11)$$

where pK is the negative logarithm of the equilibrium constant, T is temperature (°C), and parentheses refer to ion activities. The equilib-

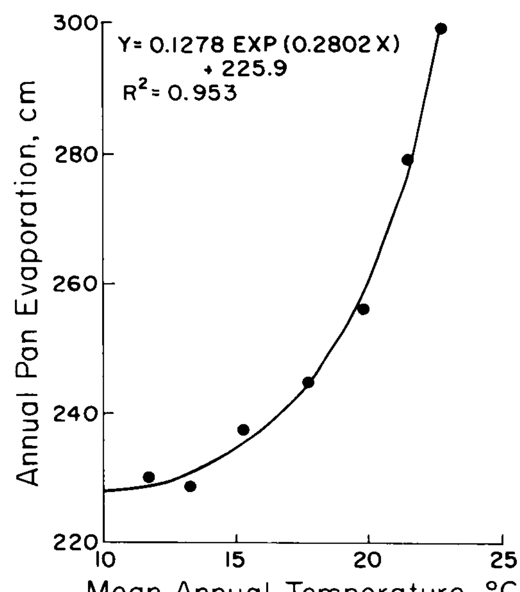

FIG. 5. Annual pan evaporation (calculated) as a function of mean annual temperature.

rium constants and their temperature dependencies were estimated from equilibrium data over the temperature range, 0 to 40°C (Garrels and Christ 1965). The intercept term in Eq. (11) was selected to yield a pK of 8.27 (25°C), which was the mean of 50 calcareous soil samples from the desert LTER site equilibrated at 25°C under a fixed CO_2 concentration (500 ppm) for 10 d. Mean monthly air temperatures were used in Eqs. (5), (7), (9), and (11).

Ionic activities (a) and concentrations (c) are related as follows

$$a = \gamma \cdot c \quad (12)$$

where γ is the activity coefficient that was estimated with the Davies equation (Sposito 1980)

$$\log \gamma = -0.505\, Z^2\left(\frac{\sqrt{I}}{1.0 + \sqrt{I}} - 0.3\, I\right) \quad (13)$$

where Z is the ionic valence, and I is the ionic strength, which was estimated by

$$I = 3.0\, c_{Ca^{2+}} \quad (14)$$

which is the theoretical relationship for a pure diunivalent salt solution (Marion and Babcock 1976). The constant (0.505) in Eq. (13) was defined for 18°C, which was the mean of the assumed range in soil temperatures (0 to 35°C). The calculated divalent activity coefficients at I = 0.1 M for the extreme temperatures (0 to 35°C) were within ±3% of the 18°C activity coefficient; this small temperature dependence of the activity coefficient was ignored in the model. For a pure $CaCO_3$ system in the pH range from 7.5 to 8.5, the following charge balance would exist

$$2\,[Ca^{2+}] = [HCO_3^-] + 2\,[CO_3^{2-}] \quad (15)$$

where brackets refer to concentrations. Substituting Eqs. (4), (6), (8), and (10) into Eq. (15) yields

$$\frac{2\,K_4(H^+)^2}{(\gamma Ca)K_3K_2K_1P_{CO_2}} = \frac{K_1K_2P_{CO_2}}{(H^+)(\gamma HCO_3)} + \frac{2K_1K_2K_3P_{CO_2}}{(H^+)^2(\gamma CO_3)} \quad (16)$$

With given CO_2 partial pressures, Eq. (16) was used to calculate the hydrogen ion activity, which has a major effect on $CaCO_3$ solubility (Garrels and Christ 1965).

The calcium equilibrium concentration was estimated from Eq. (10), after solving for HCO_3 and CO_3 concentrations, using the CO_2 partial pressures and the calculated H activity. Because the model was designed for long-term simulations, simplifications were imperative; therefore, processes such as ion-pairing, ion-exchange, and gypsum solubility, which could be important in some soils, were not included in the model. Ion-pairs between calcium and bicarbonate and carbonate are generally minor in the pH range of most calcareous soils (Marion and Babcock 1977; Suarez 1982). It was assumed that a steady-state distribution of ions existed between the solutions and exchanger phases: therefore ion-exchange was not considered in the model. The thermodynamic model is a $CaCO_3$ solubility model. The model is limited to cases where sulfate concentrations are low because: (1) sulfate ions form strong ion-pairs with calcium, (2) gypsum solubility was ignored, and (3) Eqs. (15) and (16) are invalid in the presence of high sulfate concentrations.

Soil parameterization

Each soil profile was separated into five horizons. Depending on the depth of $CaCO_3$ leaching, the five soil horizons were either 12 or 20 cm thick. Carbon dioxide partial pressures at the midpoint of each soil horizon were estimated from data published by Parada et al. (1983) for a desert soil near Tucson (Fig. 6); an assumed CO_2 partial pressure of 0.035% was used at the surface. Separate relationships were used for the winter season (November to February) and for

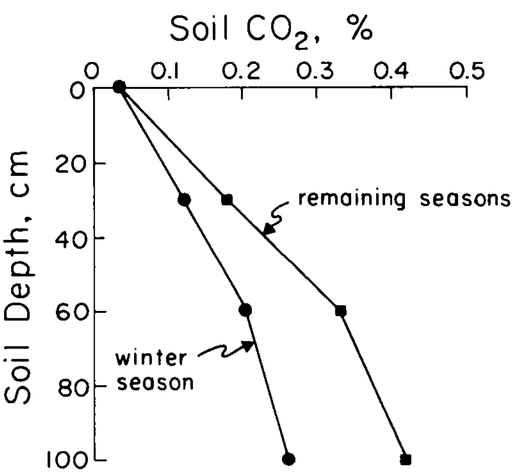

FIG. 6. Soil CO_2 concentrations as a function of soil depth (*adapted from Parada et al. 1983*).

the remaining seasons when the CO_2 partial pressure was significantly higher. We assumed for each horizon that the initial bulk density was 1.44 g/cm^3, and the initial water contents at 0.01 (field capacity) and 1.5 (permanent wilting point) MPa were 12.2 and 3.9%/weight, respectively; these data were the means of five replicates of the A horizon at the Desert LTER site near Las Cruces (Schlesinger et al. 1986). The soil WHC, which is the difference in water contents between 0.01 and 1.5 MPa, was a critical parameter in the simulation model. The consequence of selecting 8.3% (12.2 to 3.9) as the initial state will be examined in the text.

Soil water and $CaCO_3$ *fluxes*

All precipitation was assumed to enter the surface horizon. Only saturated flow through the soil profile was considered. If precipitation exceeded the WHC of the first horizon, water moved into the second horizon. This procedure continued with deeper horizons until the soil absorbed all the precipitation or the bottom horizon was reached. Water flux past the soil profile base was treated as leachate and was assumed lost from the system. Evapotranspiration losses were calculated using the previously discussed evapotranspiration model; water was first extracted from the surface horizon and then from progressively deeper horizons. After water equilibrium was established in each horizon, chemical equilibrium was reestablished between the solid phase $CaCO_3$ and the solution phase calcium.

The rate of $CaCO_3$ deposition in soils is largely controlled by the influx of calcium. Calcium can enter soils through weathering from calcareous and noncalcareous parent materials and from atmospheric deposition in dust and precipitation. Most studies in southwestern deserts suggest that atmospheric calcium influxes are sufficient to explain the rates of $CaCO_3$ deposition in noncalcareous parent materials (Brown 1956; Reeves 1970; Gardner 1972; Gile et al. 1981; McFadden 1982; Schlesinger 1985). As currently structured, the CALDEP program does not consider weathering of noncalcareous rocks, but the program can handle initial $CaCO_3$ in the soil profile and subsequent weathering of this calcareous material. The atmospheric dust $CaCO_3$ input was 0.51 g/m^2/yr, and the precipitation calcium concentration was 3.0 mg/L (Gile et al. 1981).

The CALDEP flowchart

The CALDEP program can be broken into three cycles (Fig. 7). A daily cycle calculated water loss from the soil profile on a daily time step. A rain event cycle calculated flow of rainwater and $CaCO_3$ through the soil profile whenever it rained. An annual cycle included both the later cycles plus the stochastic precipitation model and the print statements.

Simulations

Simulations evaluated the climatic, parent material, biotic, and time state factors; only the topographic state factor was not explicitly evaluated.

The CALDEP model was run using present climatic (temperature and precipitation) data assuming no $CaCO_3$ in the initial soil profile. The objective was to see at what depth $CaCO_3$ will precipitate under current climatic conditions in a uniform, noncalcareous parent material.

Three Pleistocene climatic scenarios were examined in the simulations. The first scenario was the Galloway (1983) cold-dry hypothesis with temperatures 10°C colder than present and precipitation 80% of present precipitation. The decreased precipitation was accomplished by eliminating 20% of the annual storms. The second scenario was the Wells (1966, 1979) cool-wet (summer) hypothesis with temperatures 5°C

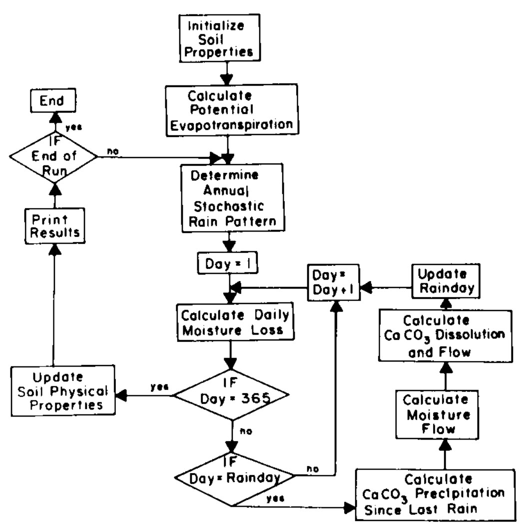

FIG. 7. A schematic diagram of the CALDEP flowchart.

colder, precipitation 67% greater than present precipitation, and the increased precipitation coming in the summer. The third scenario was the cool-wet (winter) hypothesis, which was similar to the previous hypothesis except that increased precipitation was assumed to come in the winter months (Van Devender and Spaulding 1979). The increased precipitation was accomplished by increasing the number of storms during the appropriate season (summer or winter); the increased storms were placed at random within the appropriate season. Only three of the sites (Yuma, Phoenix, and Tucson) have mean monthly temperatures > 10°C for every month; only these sites were used in the Pleistocene simulations, because other sites would have experienced freezing and thawing for substantial periods, which the model was not structured to handle.

Simulations were also run varying the soil CO_2 concentrations, WHC, and evapotranspiration rates, in order to evaluate the role of parent material and biotic factors on soil $CaCO_3$ deposition.

Validation

For validation purposes, field data from 16 Arizona soil profiles were selected (Soil Conservation Service 1974). The mean depth of the first $CaCO_3$ bulge in the soil profile was the criterion used to compare the field data and the simulation model; this mean depth, rather than the surface of the $CaCO_3$ horizon (Arkley 1963; Jenny 1980), was used, because the $CaCO_3$ surface was difficult to ascertain from the 12- to 20-cm width of the soil horizons used in the simulation model. By definition the mean depth was the weighted mean depth of the dominant horizon plus the two surrounding horizons. It being necessary that the selected soil profiles represent stable soil surfaces, soil profiles with $CaCO_3$ horizons very near the surface were not considered for fear that these profiles might represent degraded soil profiles; also soil profiles with the first $CaCO_3$ horizon greater than 1 m were not considered for fear that these profiles might represent buried soil horizons.

The soil profiles chosen in this study (Fig. 8)

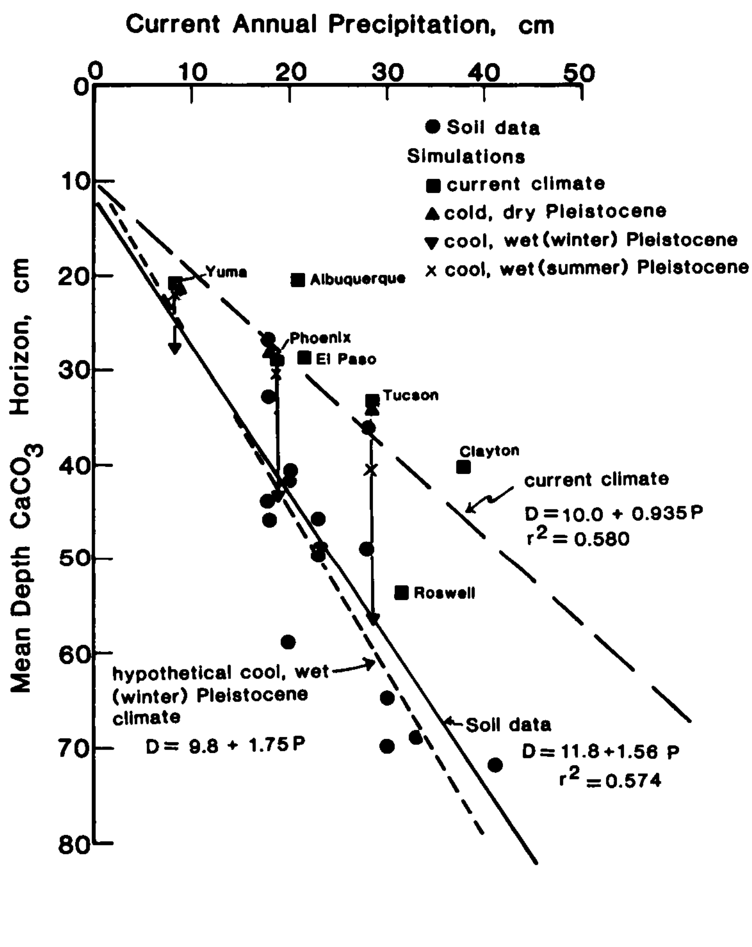

FIG. 8. The mean depth of the $CaCO_3$ horizon as a function of current annual precipitation.

were selected because they exhibited a strong $CaCO_3$ bulge in the upper soil profile; this criterion naturally selected for older profiles, for profiles younger than 10 000 yr would not have had time to develop significant $CaCO_3$ horizons. For example, the $CaCO_3$ content of the surface 1 to 2 m varied from 80 000 to 554 000 g $CaCO_3$/m^2. Assuming an average $CaCO_3$ input of 2 g/m^2/yr lead to a range of ages from 40 000 to 277 000 yr. Such profiles probably passed through several cycles of $CaCO_3$ deposition, which can be seen in multiple $CaCO_3$ bulges in many of them. If only the upper horizons including the first $CaCO_3$ bulge were considered, then the $CaCO_3$ contents ranged from 22 000 to 218 000 g/m^2; dividing by a $CaCO_3$ input of 2 g/m^2/yr lead to a range of ages from 11 000 to 109 000 years. It is clear that the upper $CaCO_3$ bulge was mostly Pleistocene in age. For this study the important question is: Can we explain the upper $CaCO_3$ bulge that mostly formed during the Pleistocene?

RESULTS AND DISCUSSION

The climatic factor

There was a strong correlation between the depth of the $CaCO_3$ horizon and current annual precipitation (Fig. 8). However, the simulation model using current climatic conditions (temperature and precipitation) predicted a shallower depth of $CaCO_3$ deposition than is observed in the field (Fig. 8); thus the field data were incompatible with the simulation model using current climatic conditions.

The Yuma, Phoenix, and Tucson sites were evaluated using the three Pleistocene climatic scenarios: cold-dry, cool-wet (winter), and cool-wet (summer). The cold-dry hypothesis had a negligible effect on the depth of $CaCO_3$ deposition (Fig. 8). Although the decreased temperature (−10°C) caused a lower evapotranspiration rate, which caused an increased leaching depth, the temperature effect was offset by the decreased precipitation (80% of present precipitation), which decreased the leaching depth. The net result was an insignificant change in the depth of the $CaCO_3$ horizon. The cool-wet (summer) hypothesis caused only a slight increase in the leaching depth (1 to 7 cm; Fig. 8), but the cool-wet (winter) Pleistocene climatic scenario caused a significant increase in the depth of $CaCO_3$ deposition for all sites (7.0 to 23.6 cm; Fig. 8). The absolute effect of the cool-wet (winter) climatic change increased with increasing precipitation. The incremental change (ΔD) in the depth of the $CaCO_3$ horizon between current and cool-wet (winter) Pleistocene climates as a function of current precipitation (P) is linear

$$\Delta D = -0.2 + 0.811\,P \qquad (r^2 = 0.992) \qquad (17)$$

Adding Eq. (17) to the current climatic equation (Fig. 8) yields

$$D = 9.8 + 1.75\,P \qquad (18)$$

Thus, the simulation model was compatible with the soil data, if we assumed a cool-wet (winter) Pleistocene climate (Fig. 8).

The cool-wet (winter) hypothesis caused a greater increase in the depth of $CaCO_3$ deposition than the cold-dry or cool-wet (summer) hypotheses, because all three factors changed in the cool-wet (winter) scenario led to an increased depth of $CaCO_3$ deposition. That is, lower temperature, higher precipitation, and a concentration of the higher precipitation in the winter months, when evapotranspiration rates were low, were factors conducive to deeper leaching in the soil profile. McFadden (1982) used a simulation model to estimate a variable called the leaching index (Li) that is directly related to $CaCO_3$ leaching depth; McFadden found that a cold-dry Pleistocene climate changed the Li by 2.4-fold, and a cool-wet Pleistocene climate changed the Li by 3.8-fold. As was true for our study, the cool-wet climatic hypothesis was more conducive to deeper leaching than the cold-dry hypothesis. The slope of the relation between current precipitation and depth to caliche, when caliche was assumed to form in Pleistocene pluvial periods, was 1.75, which was similar to that reported by Arkley (1963) for California and Nevada soils of 1.63 and to the slope of 1.56 for the Arizona soils (Fig. 8). The excellent agreement between the two soil data sets and the simulation model strongly supports the validity of the simulation model. The intercept difference between the Arkley paper (−1.9) and this paper (11.8) was probably due to our use of the mean depth instead of the surface of the $CaCO_3$ horizon as the reference point.

Gile et al. (1966) argued that $CaCO_3$ horizons in many soil profiles of the desert Southwest suggest more than one major cycle of carbonate accumulation. Nettleton et al. (1975) presented evidence that the present climate was inadequate to explain the rates of clay formation typical of older soil profiles; rather, they argued that Argids were developed during the wetter Pleistocene climates. McFadden (1982) presented evidence based on $CaCO_3$ horizon development, clay mineralogy, and iron oxyhydroxide composition, which suggested a polygenetic history of soil development in this region. The CALDEP simulation model as currently structured can account for the polygenetic development of soil profiles in southwestern deserts provided the cool-wet (winter) hypothesis was used to characterize Pleistocene climates; neither the cold-dry nor the cool-wet (summer) hypotheses for Pleistocene climates were adequate for this purpose. The excellent agreement between the soil profile data and the CALDEP simulation model using the cool-wet (winter) Pleistocene scenario supports the hypothesis that, during the late Pleistocene, the climate in the desert Southwest was cooler with wetter winters than at present.

Although the general trend was toward a greater depth of $CaCO_3$ deposition with increasing precipitation, considerable variation existed about the regression line (Fig. 8). For example, the predicted depths of the caliche horizon for Yuma and Albuquerque were similar (21 cm), even though the annual precipitation rates for the two sites were not similar (8.5 and 21.1 cm, respectively) (Fig. 8, Table 1). For the winter rainfall season, when the majority of the leaching takes place in these sites, the frequency of precipitation in excess of 1.5 cm/d was 1.4% for Albuquerque (maximum = 2.6 cm/d) and 3.8% for Yuma (maximum = 4.4 cm/d); the Yuma site was characterized by more intense, but less frequent, precipitation than the Albuquerque site. The Roswell and Clayton sites have 32 and 38 cm of annual precipitation, which fell primarily in the summer (Table 1); the mean depths of the calcic horizon for these two sites were 54 and 40 cm, respectively (Fig. 8). The Roswell site received 9.9% of its daily summer precipitation at rates greater than 3.0 cm/d (maximum = 10.7 cm/d), and the Clayton site received 6.7% of its summer precipitation at rates greater than 3.0 cm/d (maximum = 5.4 cm/d). These two examples demonstrate that the model was highly sensitive to the frequency of extreme precipitation events; these extreme precipitation events have a marked impact on the depth of $CaCO_3$ deposition. To define the precipitation regimes for specific sites requires an accurate record of the frequency of extreme precipitation events. Previous workers have suggested that 30 yr of precipitation data are needed to develop stable frequency distributions (Fogel 1981). The 4-yr record used to develop the stochastic precipitation model in this study showed regional trends correctly, but it was probably not adequate for site-specific cases.

In addition to the frequency of extreme events within given years, the frequency of extreme years also plays a role in $CaCO_3$ deposition. For example, the mean annual precipitation for Tucson is 28 cm, but for about 16% of the years, the precipitation was greater than 38 cm (1 standard deviation higher; Table 2); these extreme wet years should be much more effective than drier years in leaching $CaCO_3$.

The Arkley, Ahmad, and McFadden models for $CaCO_3$ deposition were not event-based models. As a consequence, these models did not address directly the importance of extreme precipitation events or years. McFadden (1982) argued that the depth of $CaCO_3$ deposition should reflect the average soil water balance, which, in turn, reflects average climatic conditions. Arkley (1963) acknowledged that wetter years were much more important than drier years on water movement through soils and advocated studying the soil water balance for as many years as feasible in model development. The results of our study suggested that extreme precipitation events and years played critical roles in defining the depth of $CaCO_3$ deposition in desert soils.

The parent material factor

Parent material affects $CaCO_3$ deposition primarily through its role as a calcium source and through its effect on the soil WHC. We did not examine the effect of parent material as a calcium source, but considered only $CaCO_3$ deposition in initially noncalcareous parent materials.

The WHC used in the climatic simulations (8.3%) was a measured value for the surface horizon of a desert soil (Schlesinger et al. 1986) that was felt to be representative of freshly deposited alluvium; this WHC should be useful for developing the general relationships among climate, parent material, biota, and time. To test the significance of WHC as a controlling factor, we ran simulations using a WHC of 4.0, 8.3, and 16.0%; these WHCs approximately correspond to soil textures of sand, sandy loam, and silt loam, respectively (Brady 1974). This range in WHC is typical for desert surface soils (Soil Conservation Service 1974). As the WHC increased from 4.0 to 16.0%, the mean depth of the $CaCO_3$ horizon decreased from 57 to 18 cm (Fig. 9); the simulation using 8.3% WHC (Fig. 9B) fell at an intermediate depth of 29 cm. It is clear that the WHC was a critical parameter in the simulation model.

The biotic factor

The biotic factor largely controls $CaCO_3$ deposition through its control of soil CO_2 concentrations and evapotranspiration. In most of the simulations, soil CO_2 concentrations changed with soil depth and season (Fig. 6). Measured CO_2 concentrations in desert soils are highly variable in both time and space; measurements range from 0.03 to 1.3% (Buyanovsky et al. 1981; Parada et al. 1983). The CO_2 concentrations in the standard simulations ranged from 0.05 to 0.40%, depending on the horizon and season (Fig. 6). A CO_2 concentration of 0.035% (Fig. 9D) is minimal, as it is unlikely that soil CO_2 concentrations ever drop below atmospheric CO_2 levels. Nonetheless, it is clear that CO_2 concentrations can, at times, get to significantly higher levels than those used in this model, although values in excess of 0.8% are rare in deserts (Parada et al. 1983). Theoretically, the solubility of $CaCO_3$ increases with increasing CO_2 concentration, resulting in a greater depth of $CaCO_3$ deposition. When the soil CO_2 concentration was 0.035% throughout the profile for all seasons, the model predicted that the calcic horizon would be 6 cm shallower than in the standard simulations (Fig. 9B,D).

One can hypothesize that the higher productivity of woodlands than deserts should lead to higher soil CO_2 concentrations, resulting in a deeper depth of $CaCO_3$ deposition. Such a CO_2 profile was not used in the Pleistocene simulations when woodlands probably existed over much of the desert Southwest. To some extent, this hypothetical CO_2 effect is offset by the coincident increase in evapotranspiration.

To assess the effect of evapotranspiration rates on $CaCO_3$ deposition, we ran the simulation model using both the Veihmeyer-Hendrickson (V-H) model, which assumed that water was lost at the potential rate across the whole available moisture range, and the Thornthwaite-Mather (T-M) model, which assumed that moisture loss was linearly related to moisture content across the same range (Fig. 4). The V-H model, which removed soil moisture rapidly from the soil, produced a mean depth of $CaCO_3$ deposition of 22.4 cm (Fig. 10); the T-M model, which removed soil moisture slowly from the soil, produced a mean depth of 32.7 cm (Fig. 10); and the model used in the standard simulations resulted in a mean depth of 28.6 cm (Fig. 9B). The range in these predictions (10.3 cm) was sufficiently broad that some caution is necessary in selecting the evapotranspiration model. For example, Specht (1972) has shown that the regression for soil moisture removal shifts from the right to the left as xerophytic vegetation is supplanted by mesophyte vegetation (Fig. 4). This factor was not considered in the Pleistocene simulations. However, based on Specht's work (1972), one can hypothesize that as desert veg-

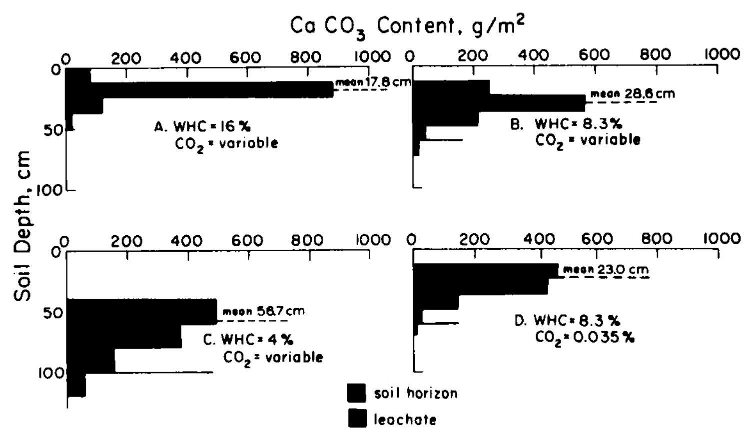

FIG. 9. The soil $CaCO_3$ content as a function of water-holding capacity (WHC) and carbon dioxide concentration for the El Paso site (500-yr simulations).

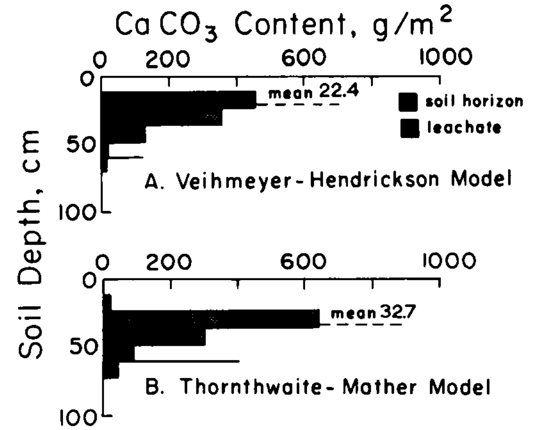

FIG. 10. The soil $CaCO_3$ content as a function of evapotranspiration rate for the El Paso site (500-yr simulations).

etation replaces woodlands, water removal rates will decrease, at a given temperature and soil moisture content, and $CaCO_3$ deposition will be deeper.

It is clear that the biotic factors that control soil CO_2 partial pressures and evapotranspiration rates were critical parameters in the simulation model; changes in these parameters with changing vegetation are particularly important in simulating long-term changes in $CaCO_3$ deposition. When vegetation shifts from desert to woodland, one can hypothesize a shallower depth of deposition due to changes in the evapotranspiration rate and a deeper depth of deposition due to changes in the CO_2 concentration. The net effect of these two opposing trends is unclear at present. Work is currently underway to develop the data base needed for evaluating the simulation model under woodland and grassland conditions.

The time factor

The different sites showed an increasing rate of $CaCO_3$ deposition with increasing precipitation. This was a consequence of the model, which assumed a constant dust input (0.5 g $CaCO_3/m^2/yr$) and a constant calcium concentration in the precipitation (3 mg/L). Therefore, as total precipitation increased, total calcium input increased. Based on current precipitation, the predicted rates of $CaCO_3$ deposition for Yuma, Albuquerque, El Paso, and Roswell were 1.2, 2.1, 2.2, and 2.9 $g/m^2/yr$, respectively. Based on the cool-wet Pleistocene precipitation, the predicted rates of $CaCO_3$ deposition for these four sites were 1.6, 3.3, 3.4, and 4.5 $g/m^2/yr$. There is evidence from field studies of an increasing $CaCO_3$ deposition rate with increasing precipitation. For Vidal Junction, California (near Yuma), Albuquerque, Las Cruces (near El Paso), and Roswell, estimates of $CaCO_3$ deposition rates were 0.95, 2.2, 3.2, and 5.1 $g/m^2/yr$, respectively (Bachman and Machette 1977). The field estimates agreed reasonably well with the corresponding simulation model estimates based on either current or Pleistocene precipitation. Although reported values for soil $CaCO_3$ deposition in southwestern deserts range from 1 to 12 $g/m^2/yr$, the majority of estimates fell within a narrower range—from 1 to 5 $g/cm^2/yr$ (Schlesinger 1985). This narrow range agreed with the simulation model, where the $CaCO_3$ deposition rates ranged from 1.2 (Yuma, current precipitation) to 5.3 (Clayton, Pleistocene precipitation) $g/m^2/yr$. Given the simplicity of the model and the numerous assumptions inherent in the field estimates, the agreement between the field data and the simulation model supports the use of the model for regional generalizations.

One of the characteristics of old $CaCO_3$ profiles is the eventual plugging of the $CaCO_3$ horizon and the subsequent formation of a laminar layer above the plugged horizon (Gile et al. 1981). The rate at which plugging occurs depends on the rate of calcium input and calcium concentration within a given horizon. The rate of $CaCO_3$ deposition within the dominant $CaCO_3$ horizon stabilized early in the simulations (Fig. 11). The slopes of the curves (Fig. 11) gave the rate at which $CaCO_3$ was accumulating within the dominant $CaCO_3$ horizon; dividing the slope by the total Ca input rate for the site

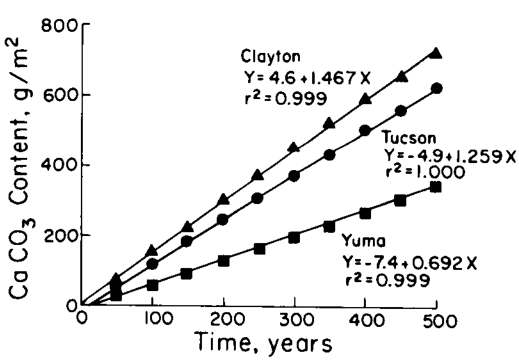

FIG. 11. The rate of $CaCO_3$ deposition in the dominant $CaCO_3$ horizon for three selected sites.

gave the relative degree of concentration of incoming calcium. For the Yuma, Tucson, and Clayton sites, the degrees of concentration were 0.60, 0.48, and 0.43, respectively; the lower the precipitation, the greater the degree of concentration and the closer to the surface the dominant $CaCO_3$ soil horizon developed. This concentration in the surface horizons can also be seen when the WHC was high (Fig. 9A). Gile et al. (1981) estimated that plugging of $CaCO_3$ horizons occurs at approximately 40% $CaCO_3$; for a 12-cm horizon with a bulk density of 1.44 g/cm^3, 40% $CaCO_3$ is equivalent to 69 100 g $CaCO_3$/m^2. Dividing the slopes of Fig. 11 into 69 100 gives an approximate time for plugging of the dominant soil horizon; for the Clayton, Tucson, and Yuma sites, the calculated times were 47 000, 55 000, and 100 000 yr, respectively, assuming current calcium inputs. Assuming the cool-wet Pleistocene climate, the calculated times for horizon plugging for the Tucson and Yuma sites were 66 000 and 101 000 yr, respectively. A greater time for plugging under a Pleistocene climate when total calcium input was substantially higher occurred because the degree of concentration of $CaCO_3$ within a horizon decreased with increasing precipitation. Gile et al. (1981) estimated that it takes a minimum of 25 000 to 75 000 yr for plugging to occur for gravelly soils near Las Cruces. These field observations agreed reasonably well with the ranges derived from the simulation model.

For long-term simulations, factors including bulk density, WHC, climate, and vegetation can potentially change the environment for $CaCO_3$ deposition. With time, the bulk density of carbonate horizons would increase as $CaCO_3$ and clays fill the interstitial spaces; the effect of $CaCO_3$ filling was incorporated in the model, but clay formation was not. The WHC should increase up to a certain age due to clay formation and then decrease as the pore space is reduced due to interstitial filling. The model did not consider a changing WHC. Climatic fluctuations were widespread in the past; the two climatic regimes (current and Pleistocene) used in the simulations were two regimes around which climate has fluctuated widely. Because it takes only about 100 to 200 yr of simulation for a stable soil profile to develop under a given climatic regime, the short-term simulations adequately describe the effect of a given climate on $CaCO_3$ deposition. However, the simulation of long-term changes would require a changing climatic regime. The effect of a changing biota on soil parameterization for long-term simulations was discussed previously.

The present program was developed using the BASIC language on a Hewlett-Packard Model 9816 minicomputer (16-bit processor) and requires approximately 75 min to simulate 100 yr; to simulate 100 000 yr would require running the program for 52 d continuously; because supercomputers are approximately 1000-fold faster than minicomputers (Stern and Stern 1982), a supercomputer would reduce the processing time to 1.25 h.

Despite its limitations for long-term simulations, the model as now structured can be used to assess the roles of climate, parent material, biota, and time on $CaCO_3$ deposition based on short-term simulations. The great utility of the CALDEP simulation model is that it allows one to evaluate the role of state factors separately; in this way, the model is a valuable research tool in ascertaining the role of state factors on $CaCO_3$ deposition in soils.

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