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RATES OF SOIL FORMATION: IMPLICATIONS FOR SOIL-LOSS TOLERANCE¹

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Land managers are concerned with maintaining crop, forest, and range production. If erosion losses from moderately deep or shallow soils occur faster than the soil is replenished by weathering from underlying bedrock or consolidated sediments, productivity will decline. Therefore, rates of soil formation can be considered constraints on tolerable soil losses from relatively shallow soils; their long-term soil-loss tolerances can be equated with their rates of formation from bedrock or consolidated sediments. There are few data for determining rates of soil formation, however. Data from small, nonagricultural watershed studies in which the inputs and outputs of silica and the major cations have been measured can be utilized for this purpose. Rates of soil formation, based on these data from 18 watersheds with noncarbonate lithologies, range from 0.02 to 1.9 Mg/ha yr. The rates are most highly dependent on (1) the volume of runoff water and (2) the soil-to-rock ratio, where this ratio is a mass of soil divided by the mass of bedrock weathered to produce that mass of soil. An equation developed to predict rates of soil formation from these two parameters has a very high coefficient of determination, $R^2 = 0.94$.

Land managers are concerned with maintaining crop, forest, and range production. Reductions in soil depth due to erosion or the mechanical displacement of surface soil reduce productivity, because the storage of plant-available water and some nutrients is a function of the mass and volume of soil. Soil losses from very deep soils may not reduce productivity if losses are so slow that the soil organic matter content can be maintained without declining. If erosion losses from moderately deep or shallow soils are faster than the mass of soil is replenished by weathering underlying bedrock or consolidated

sediments, however, productivity will decline, even though reductions may not be measurable over short periods. Therefore, rates of soil formation can be considered constraints on tolerable soil losses from shallow and moderately deep soils. The tolerable soil losses for these soils can be equated to their rates of formation from bedrock or consolidated sediments (Stocking 1978).

Rates of soil formation must be determinable or predictable before they can be used as criteria for setting soil-loss-tolerance limits. Just one-quarter of a century ago, little was known about the rates at which soils are formed from bedrock or consolidated sediments. Smith and Stamey (1965), in a thorough review of the literature, did not find any reliable estimates of soil formation from consolidated materials, and they may not have missed any pertinent articles other than Anderson's (1941), which I missed too (Alexander 1985). Recent review articles on soil-loss tolerance in relation to rates of soil formation dwell on soil horizon differentiation, rather than on soil formation from lithic and paralithic materials (Hall et al. 1985), even though the latter may be more pertinent (Stocking 1978). The common emphasis on soil horizon differentiation may be due less to its presumed primary importance than to the lack of quantitative information on rates of soil formation from bedrock or consolidated sediments.

The most reliable estimates of rates of soil formation from consolidated sediments or bedrock are based on small watershed studies, which have become much more numerous in the last 20 years. Data from these studies have been reviewed and rates of soil formation computed from an elemental balance equation of Barth (1961), where rock and soil analyses were available, or from a mass balance approximation of Barth's equation where only runoff quantity and chemistry were available (Alexander 1985). The approximate equation is

$$W = D + S \quad (1)$$

where W is the mass of rock or consolidated sediments weathered, S is the mass of residue

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or soil formed from the weathered lithic or paralithic material, and D is the mass of dissolved solids removed in runoff from the soil-parent material system. It is approximate because chemical elements in the soil may have oxidation states different from those of the same elements in its parent material, and thus different masses of associated oxygen. The approximate equation, however, is reasonably accurate in most cases. Because both W and S are unknown variables, the rate of weathering is computed first by a transformation of Barth's equation or the approximate equation containing an S/W ratio rather than S

$$W = D/(1 - S/W) \quad (2)$$

then $S = W(S/W)$. The S/W ratio must be estimated for solving the approximate equation, but it can be done with considerable confidence.

There are still too few reliable calculations or estimates of rates of soil formation for developing a model to predict accurate rates for soil formation from specific parent materials on specific landforms in specific climates. However, I have developed an equation that can be used to define the ranges of values that might be expected for rates of soil formation from common noncarbonate rocks or consolidated deposits. These ranges may also be considered the ranges of the soil-loss tolerance for shallow or moderately deep soils.

MODEL

Chemical erosion, or denudation, and soil formation are closely related to the volume of runoff from small watersheds (Alexander 1985). Hence, I have developed a model to predict rates of soil formation based on runoff. It is a crude model, because many factors unrelated to runoff affect the rates of soil formation. Nevertheless, with the close relationship to runoff, the model is expected to yield good first approximations of rates of soil formation from bedrock or consolidated sediments. "The term runoff is usually considered synonymous with streamflow and is the sum of surface runoff and groundwater flow that reaches the streams" (Davis and DeWiest 1966).

The basis for my development of equations for rates of chemical denudation and soil formation is a relationship used by Langbein and Dawdy (1964) to develop an equation for dis-

solved load, or the concentration of dissolved solids, in stream water

$$dC/dt = k(C_m - C)/C_m \quad (3)$$

where C is the actual concentration and C_m is the maximum (or equilibrium) concentration of dissolved solids, dC/dt is the rate of dissolution of the lithic or paralithic material, and k is a constant. Integrating, and assuming that C is negligible when $t = 0$

$$\ln(1 - C/C_m) = -kt/C_m \quad (4)$$

The time that a molecular of water remains in the soil or in the proximity of the lithic or paralithic contact is inversely proportional to the amount of precipitation, infiltration, and hence runoff, for runoff is generally sustained by the displacement of subsurface (stored) water (Sklash and Farvolden 1982); that is $Q = \text{constant}/t$, where Q is the amount of runoff in time, t . This is a gross simplification, but it should be acceptable for a model used simply to provide a rational equation for regression analysis, rather than an exact process simulation. Chemical denudation is equal to the product of runoff and concentration; $D = QC$. The substitution of $t = \text{constant}/Q$ and $C = D/Q$ in Eq. (4) leads to

$$D = CmQ(1 - e^{-k'/CmQ}) \quad (5)$$

Alternatively, Eq. (5) can be derived from an analysis of empirical data to show that it is not an accidental result from faulty assumptions in the deductive derivation. A plot of total dissolved solids as a function of annual runoff (Stumm and Morgan 1981, p. 535) shows that the concentration (C) is nearly constant or declines slightly as runoff increases from zero to 2 cm/yr and then decreases substantially with further increases in runoff. This response of solute concentration to runoff implies that the concentration declines from a maximum (C_m) when there is no runoff ($Q = 0$) at a rate proportional to the departure of concentration from the maximum; that is, $-dC/dQ = k(C_m - C)/C_m$. As runoff becomes very great, both C and its decrease in relation to Q approach zero. This extreme can be represented by $-dC/dQ = \text{constant}/Q^{n+1}$, where $n + 1$ is chosen as the exponent of Q rather than n to simplify the result of integration. Combining these two functions, $-dC/dQ = k'(1 - C/C_m)/Q^{n+1}$, and integrating, $C_m \ln(1 - C/C_m) = k'Q^{-n}/(-n) + K$, where K is the constant of integration. As Q approaches

infinity, C approaches zero; then, when $C = 0$, $K = 0$. Eliminating K and solving for solute concentration, $C = Cm(1 - e^{-k'/CmQ})$ when $n = 1$. Since $D = QC$, $D = CmQ(1 - e^{-k'/CmQ})$, which is the same as Eq. (5).

To obtain an equation for rates of soil formation, I transformed the mass balance equation (Eq. (1)) for the weathering of rock (W) to soil (S) and dissolved solids (D) to obtain $D = S(W/S - 1)$, where W/S is the inverse of the soil-to-weathered-bedrock (S/W) ratio discussed by Alexander (1985). Substituting this relationship for D into Eq. (5) leads to

$$S = CmQ(1 - e^{-k'/CmQ})/(W/S - 1) \quad (6)$$

DATA AND METHODS

Published data from 18 small watersheds (Table 1) were used to develop the constant and coefficients for the model. To obtain data from areas of predominantly one lithology, only small watersheds were selected. Watersheds of predominantly sedimentary rocks were avoided except for graywacke in Wales. Thus, because the concentration was on rocks with negligible amounts of anions other than oxygen, the chemical losses from most watersheds were essentially determinable by summing the major cations (Na, K, Ca, Mg) and silica (SiO_2). Some watersheds, however, have appreciable losses of Fe and Al (Reid et al. 1981). Bicarbonate is usually the major anion in runoff water, but it is derived directly or indirectly from atmospheric CO_2 . By excluding watersheds with dominant lithologies containing carbonates, the bicarbonate in runoff could be ignored for calculating chemical denudation. The Victoria, Australia, watersheds with dacite (Table 1) were not included in the regression analyses, because volcanic rocks commonly weather more rapidly than plutonic and noncarbonate metamorphic rocks.

Soil and parent material analyses were available for 7 of the 18 watersheds with data used in regression analyses. For the Scotland watershed (SG), for example, the mean concentrations of Al and Fe in till (W), in combined E and B horizons (S), and annual losses in drainage water (D) were substituted into Barth's elemental balance equation (Alexander 1985) to obtain two equations, one for Al and another for Fe; then these two equations were solved simultaneously to obtain W and S . The 11 watersheds

without soil and parent material analyses are distinguished in Table 1 by S/W ratios in parentheses, because the S/W ratios were estimated rather than computed for these 11 watersheds. These estimates were developed from the data and S/W relationships discussed by Alexander (1985). They are based on parent material and soil development. The S/W estimates are 0.95 for shallow soils ($n = 0$), 0.9 for watersheds with Inceptisols and shallow soils ($n = 1$), 0.85 for moderately deep and deep Inceptisols and Spodosols in till ($n = 7$), 0.8 for Alfisols ($n = 2$), and 0.7 for Ultisols ($n = 1$). Although the weathering of Alfisols may be just as great as that of Ultisols in low-latitude, mountainous areas (for example, Zimbabwe, Table 1, ZJ and ZR), the weathering of Ultisols (MS, Table 1) is generally greater than that of Alfisols (MB and MU, Table 1) at low altitudes in continental climates.

Regression analyses were applied to determine the constants in Eqs. (5) and (6) and coefficients. The constants k' in Eqs. (5) and (6) were chosen such that the exponents of some or all of the terms would be one.

Correlations were determined between the variables in Eq. (6) and watershed latitude (degrees) and relief ratio. Whereas the relief reported in Table 1 is simply the elevation difference from the highest to the lowest points in a watershed, the relief ratio is the relief (Table 1) divided by the square root of the watershed area ($\text{m}/\text{ha}^{0.5}$ or $\text{m}/100 \text{ m}$). Because the relief ratio depends on the watershed area, drainage density would likely have been a more satisfactory variable, if that datum had been available.

RESULTS AND DISCUSSION

The logarithmic correlations of runoff with both chemical denudation and soil formation are very highly significant (Table 2), the correlation being stronger for chemical denudation. Also, chemical denudation is strongly correlated (logarithmically) with relief and the S/W ratio, and the rate of soil formation is strongly correlated (logarithmically) with latitude. The correlation of the rate of soil formation with the S/W ratio is an expectable result, for the S/W ratio is used in computing the rate of soil formation. The S/W ratio is strongly correlated with latitude too. It is a positive correlation, because chemical weathering of the soil parent material is less complete in the cooler climates of higher latitudes than at lower latitudes. Because soils

TABLE 1
Annual rates of chemical denudation (D) and soil formation (S) in small watersheds of predominantly one lithology

Sym. ^a	State or country	Lat., deg.	Area, ha	Relief, m	Lithology	Soils	Vegetation	Run-off, cm ^b	D, kg/ha	S/W ^b	Calculated Mg/ha	S Pred. Eq. (10)
CA	Calif.	37	274	220	Adamellite	Cryoborolls, Cryochrepts	Sagebrush	14	54	0.70	0.13	0.09
CH ^c	BC, Can.	49	47	185	Till/quartz diorite	Podzols or Haplorthods	Conifer forest	151	127	(0.85)	0.72	0.69
CS	BC, Can.	50	300	945	Till/quartz diorite	Podzols	Conifer forest	367	164	(0.85)	0.93	1.00
EG	England	51	475	232	Granite	Ocherts, Orthods	Spruce forest, grassland, bog	(130)	116	(0.85)	0.66	0.65
FR	France	43	146	152	Gneiss	Rankers	Maquis	65	67	0.95	1.27	1.60
IG	Idaho	44	144	478	Adamellite	Orthents, Umbrepts	Conifer forest	39	107	0.88	0.78	0.49
LB	Luxemb.	50	17	90	Metashale	Dystrochrepts, Lithic	Deciduous forest	38	61	(0.90)	0.55	0.59
MB	Maryland	39	550	280	Greenstone (meta-bas.)	Udortheints	Deciduous forest	55	109	(0.80)	0.44	0.31
MG	Minn.	48	2520	98	Till/gabbro	Orthods(?) and peat	Mixed forest, spruce bog	27	42	(0.85)	0.24	0.31
MS	Maryland	39	38	67	Schist	Hapludults	Deciduous forest	16	31	(0.70)	0.07	0.10
MU	Maryland	39	57	70	Serpentine	Hapludalfs	Open pine-oak forest	14	59	(0.80)	0.24	0.15
NH	N. Hamps.	44	28	300	Till/gneiss	Haplorthods	Mixed forest	71	88	0.80	0.35	0.35
SG	Scotland	57	4120	542	Till/granite	Peat and Podzols	Heather moorland	118	121	0.94	1.90	1.72
V1	Victoria	37S	25	144	Dacite	Kraznozems	Eucalyptus forest	70	153	0.72	0.39	0.22
V2	Victoria	37S	30	203	Dacite	Kraznozems	Eucalyptus forest	90	264	0.72	0.68	0.25

WA	Wash.	47	520	1150	Till/quartz diorite	Inceptisols and Spodo- sols	Subalpine	440	180	(0.85)	1.02	0.71
WF	Wales	53	7	105	Till/wacke	Acid Brown Earths	Oak forest	31	54	(0.85)	0.31	0.34
WP	Wales	53	10	90	Till/wacke	Acid Brown Earths	Pasture	30	40	(0.85)	0.23	0.33
ZJ	Zimbabwe	18S	91	160	Adamellite	Hapludults	Acacia sa- vanna	40	74	0.75	0.22	0.20
ZR	Zimbabwe	18S	733	60	Adamellite	Paleustalfs	Grassland	8	19	0.57	0.02	0.03

^a References: CA, Marchand (1971, 74); CH, Feller and Kimmings (1979); CS, Zeman (1975); EG, Williams et al. (1988); FR, Martin (1983); IG, Clayton and Megahan (1986); LB, Verstraten (1977); MB, Katz et al. (1985); MG, Siegel and Pfannkuch (1984); MS, Cleaves et al. (1970); MU, Cleaves et al. (1974); NH, Likens et al. (1977); SG, Reid et al. (1981); V1, V2, Anderson (1941), Feller (1981); WA, Dethier (1979); WF, WP, Oxley (1974); ZJ, ZR, Owens and Watson (1979, 79).

^b Numbers in parentheses are estimated, rather than given in the references cited or calculated.

with higher *S/W* ratios are commonly either relatively shallow or in cooler climates, the positive logarithmic correlation of runoff with the *S/W* ratio may be due to lower productivity and hence less evapotranspiration and more runoff from areas with higher *S/W* ratios.

It may seem surprising that the rate of soil formation is positively correlated with latitude. This positive correlation reflects the greater influence of the *S/W* ratio, which is positively correlated with latitude, than the rate of weathering or chemical denudation, which is not correlated significantly with latitude, on the rate of soil formation. Soil formation is probably most rapid for shallow or rudimentary soils at lower latitudes. The two Zimbabwe watersheds at relatively low latitudes (18°S), however, contain deep, highly weathered soils.

The regression equation for chemical denudation corresponding to Eq. (5) is

$$D = 2.56Q(1 - e^{-55/Q}) \quad r^2 = 0.762 \quad (7)$$

If the lone *Q* term is treated as a separate independent variable, the regression equation becomes

$$D = 1.79Q^{1.34}(1 - e^{-13.3/Q}) \quad R^2 = 0.816 \quad (8)$$

where runoff (*Q*) is in cm/yr and chemical denudation (*D*) is in kg/ha yr. Lines representing each of these two equations are reproduced in Fig. 1. They are very close to one another with runoff <200 cm/yr, but they diverge considerably with greater runoff.

The regression equation for rates of soil formation corresponding to Eq. (6) is

$$S = \frac{0.00277Q(1 - e^{-48/Q})}{W/S - 1} \quad r^2 = 0.925 \quad (9)$$

If the lone *Q* term is treated as an independent variable, the regression equation becomes

$$S = \frac{0.00243Q^{1.40}(1 - e^{-6.9/Q})}{W/S - 1} \quad (10)$$

$$R^2 = 0.942$$

Lines representing Eq. (9) are reproduced in Fig. 2. Rates of soil formation predicted by either equation are nearly the same where the runoff is less than 200 cm/yr. Divergence with greater runoff is similar to that for Eqs. (7) and (8) predicting chemical denudation.

Although the maximum concentration of dissolved solids, *C_m*, is a constant for a particular

TABLE 2

Correlation (r) of variables based on data from 18 watersheds; linear coefficients are tabulated on the upper right, and logarithmic coefficients are on the lower left^a

	Chemical denudat. D	Soil format. S	Annual runoff Q	Relief ratio RR	Soil/rock S/W	Latitude L	
$\log D$	-	0.65*	0.87**	0.56	0.46	0.38	D
$\log S$	0.85**	-	0.50	0.16	0.73**	0.51	S
$\log Q$	0.89**	0.80**	-	0.59	0.28	0.31	Q
$\log RR$	0.60*	0.48	0.55	-	0.27	0.33	RR
$\log S/W$	0.62*	0.91**	0.60*	0.41	-	0.78**	S/W
$\log L$	0.45	0.65*	0.48	0.39	0.77**	-	L
	$\log D$	$\log S$	$\log Q$	$\log RR$	$\log S/W$	$\log L$	

^a Highly or very highly significant correlation coefficients are indicated by * ($p < 0.01$) or ** ($p < 0.001$).

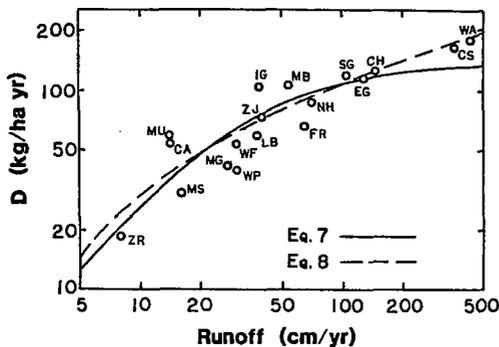


FIG. 1. Chemical denudation (D) from 18 small watersheds with noncarbonate lithologies.

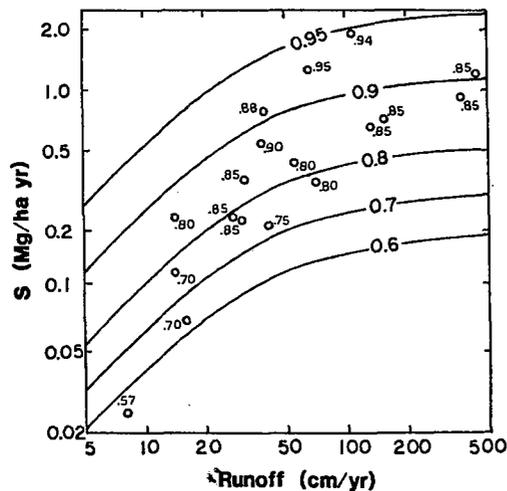


FIG. 2. Rates of soil formation (S) in 18 small watersheds with noncarbonate lithologies. The curves were determined by substituting $S/W = 0.6, 0.7, 0.8, 0.9,$ and 0.95 into Eq. (9).

site, it may vary from one watershed to another. It depends on the parent lithology, temperature, and CO_2 supply. Using latitude as a surrogate for temperature, I substituted it for C_m in Eqs. (8) and (10). The coefficients of determination increased only slightly, from 0.816 to 0.823 for Eq. (8) and from 0.942 to 0.943 for Eq. (10).

Rates of soil formation predicted by Eq. (10) are close to computed values (Fig. 3), except for the watersheds of Idaho (IG) and France (FR) at $S = 0.785$ and 1.273 Mg/ha yr, respectively. The relatively poor predictions for these two watersheds are more evident on linear scales; a logarithmic scale was used, in Fig. 3, however, to avoid congestion at lower rates of soil formation, because the rates are < 0.5 Mg/ha yr in 10 of the 18 watersheds. Rates predicted by Eq. (9) are practically as good as those of Eq. (10), except for the Seymour, B.C., and Washington

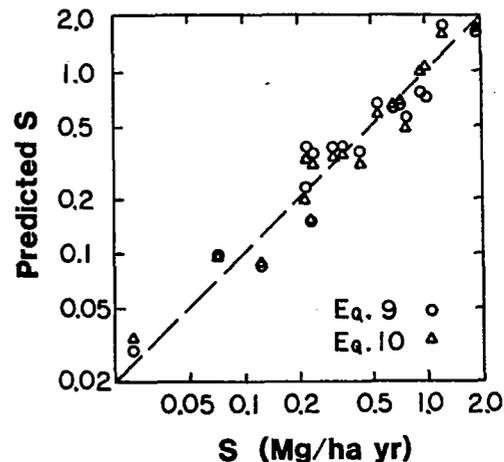


FIG. 3. A comparison of calculated (Barth's equation or Eq. (2)) and predicted (Eq. (9), circles; Eq. (10), triangles) rates of soil formation.

watersheds at $S = 0.929$ and 1.020 Mg/ha yr, respectively. These are the only watersheds with runoff >200 cm/yr, indicating that Eq. (10) may be more reliable than Eq. (9) for projections beyond 200 cm of runoff.

The equations developed from watersheds with noncarbonate lithologies, mostly igneous and metamorphic rocks, underestimate the rates of chemical denudation and soil formation in many watersheds with predominantly sedimentary rocks. Two watersheds in Virginia, one with sandstone and the other with calcareous shale, in which the rates of soil formation were 0.22 to 1.00 Mg/ha yr (Alexander 1985), were predicted by Eq. (10) to have rates of 0.31 and 0.27 Mg/ha yr, respectively. A watershed in England with boulder clay over calcareous shale and some sandstone (Foster and Grieve 1984) had an estimated rate of soil formation of 2.65 Mg/ha yr compared with a predicted rate of only 0.27 Mg/ha yr, assuming an S/W ratio of 0.85 . Rates of soil formation are greatest where sedimentary rocks with resistant grains are bound in a readily weatherable cementing material. Equations (9) and (10) are not applicable to sedimentary or metamorphic rocks with appreciable carbonates, chlorides, sulfates, or other relatively soluble compounds.

Many stream data are available from the Coweeta Hydrologic Laboratory, North Carolina. Mature oak-hickory-maple forest has been retained in seven of the Coweeta watersheds. Runoff from these undisturbed watersheds ranges from 93 to 183 cm/yr (Swank and Waide 1987). Chemical denudation increases with annual runoff from the four watersheds where the Tallulah Falls Formation of schists and other metamorphic rocks predominates (Table 3). It is less from watersheds of comparable runoff where the Coweeta Group of gneisses and other metamorphic rocks predominate. Predictions of chemical

denudation (Eqs. (7) and (8)) and soil formation (Eqs. (9) and (10)) rates are higher than the calculated rates based on the sum of cations plus silica and an S/W ratio of 0.6 (Table 3). The paired t test indicates that they are significantly different ($p < 0.05$) for both D and S for watersheds of Coweeta Group terrain, but that only predictions of Eqs. (8) and (10) are significantly different from computed values for watersheds of the Tallulah Falls Formation. Predictions of chemical denudation by Eq. (7) and soil formation by Eq. (9) are not significantly different from computed values for watersheds of the Tallulah Falls Formation. This does not mean that Eqs. (7) and (9) are more reliable than Eqs. (8) and (10), but merely that the means of predictions by Eqs. (7) and (9) are closer to the means of calculated rates of chemical denudation and soil formation for these particular watersheds. It must be remembered that these are general equations developed from data for watersheds of many lithologies; more accurate equations can be developed for specific lithologies when more data become available.

Velbel (1985) using cation balance equations involving the weathering of biotite, almandine, and oligoclase (Coweeta Group) or andesine (Tallulah Falls Formation), and biotic uptake, computed saprolization rates at the Coweeta Hydrologic Laboratory. Assuming isovolumetric weathering from bedrock to saprolite having a bulk density of 1.5 Mg/m³, he computed a saprolization rate of 3.8 cm/1000 yr for Watershed 27 where rocks of the Coweeta Group predominate. A saprolite bulk density of 1.5 Mg/m³ indicates an S/W ratio slightly less than 0.6 . Assuming it is 0.6 , I computed a saprolization rate of 0.78 cm/1000 yr, based on a chemical denudation rate of 77.6 kg/ha yr from Watershed 27 (Swank and Waide 1987). The highest rate I computed was 1.29 cm/1000 yr (Wa-

TABLE 3

Rates of chemical denudation (D) and saprolite formation (S) in undisturbed watersheds of the Coweeta Hydrologic Laboratory where rocks of the Tallulah Formation predominate over those of the Coweeta Group

Watershed	Runoff Q, cm/yr	D	Predicted D		S	Predicted S	
			Eq. (7) kg/ha yr	Eq. (8)		Eq. (9) Mg/ha yr	Eq. (10)
2	93	92.0	106.3	103.6	0.138	0.155	0.149
18	119	98.4	112.7	114.4	0.147	0.164	0.165
34	128	98.3	114.5	117.7	0.147	0.166	0.171
36	183	129.2	121.6	135.0	0.193	0.175	0.198

tershed 36). Velbel's estimate for Watershed 27 appears to be too high, possibly by a factor of 5. His method, based on the fluxes of four cations, may be less accurate for determining the saprolization rate than the method I used based on Barth's equation, because 84% of the chemical denudation from Watershed 27 is due to SiO_2 losses.

CONCLUSIONS

Assuming that the soil-loss tolerance of shallow and moderately deep soils can be equated to their rates of soil formation, the tolerances for these soils forming from plutonic or noncarbonate metamorphic rocks can be estimated by substituting runoff and an S/W ratio into Eqs. (9) or (10). Only 1% of the area in the 48 contiguous states has runoff >100 cm/yr (Langbein 1949). When 100 is substituted for Q in Eq. (10), the predicted values of soil formation (S) range from 0.3 Mg/ha yr for $S/W = 0.75$ through 0.9 for $S/W = 0.9$ to 1.9 Mg/ha yr for $S/W = 0.95$. Soil-loss tolerances for soils formed from calcareous sandstones and shales may be higher, but most shallow and moderately deep soils formed from plutonic rocks should have tolerances much lower than 1 t/A yr (2.24 Mg/ha yr).

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