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The Physico-Chemical Properties of Soils Affecting Soil Erosion

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ABSTRACT

A study of the physical properties influencing the erosiveness of the Iredell sandy clay loam and the non-erosiveness of the Davidson clay in North Carolina showed that the Davidson was highly granulated into large, stable and porous aggregates in contrast to the lower content of small, compact granules, and the dispersed condition of the Iredell. Physico-chemical studies of the extracted colloids from the Iredell, Davidson, and Putnam soils showed that the electro-dialyzed Davidson colloids flocculated when saturated with H-, Li-, Na-, K-, Ca-, or Ba-ions, or, with Na-ions up to 150 per cent saturation, whereas only the H-, Ba-, and Ca- systems of the Iredell and Putnam colloids were flocculated.

The zeta-potentials of the three colloids were about the same, indicating that the charge of the particles is probably a less important factor than hydration influencing the stability of colloidal clay suspensions. Viscosity measurements indicated that the dispersed systems of the Putnam were more hydrated than the Iredell. Swelling determinations showed that the order of hydration was bentonite > Putnam > Iredell > Davidson. Permeability of all colloids was in the order: H > Ba > Ca > K > Na > Li, which for the H-, Ba-, and Ca- systems is the same as the order of viscosity, but the reverse of swelling. Data indicate that the permeability of the soil profile and the ease of dispersion of the soil particles are extremely important factors affecting soil erosion.

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The Physico-Chemical Properties of Soils Affecting Soil Erosion*

J. FULTON LUTZ†

INTRODUCTION

Soil erosion is one of the greatest contributing factors to soil deterioration. A study of erosion losses by Bennett and Chapline^{1,2,‡} has shown that 35,000,000 acres of farm land have been essentially destroyed by man-induced erosion and that the present destruction is at the rate of more than 100,000 acres a year. The severity of these losses requires careful study of the various factors affecting the erosion of different soils in order to accumulate sufficient facts upon which an efficient method of soil conservation can be based.

It is recognized that erosion is influenced by a number of meteorological and environmental factors, such as the amount and intensity of rainfall, the slope and area of the land^{2,3}, the nature and amount of vegetation, the amount of moisture in the soil at the time of rain, and other related influences. Some of these factors can in part be directly controlled by suitable soil management practices; others can only be influenced by indirect means. Experiments have shown that in humid regions a rotation system of farming which includes a legume as an erosion-preventing crop will decrease surface run-off and consequently diminish erosion losses^{5 18 32}. Reforesting the steeper slopes or seeding them to pasture grasses effectively controls erosion. The amount and intensity of the rainfall can not be controlled although the detrimental effects can be minimized by reducing the rate of run-off with terraces.

It is generally observed that some soils erode easily while others under the same conditions of rainfall, vegetation and topography erode very little. This suggests that the nature of the soil plays an important role in erosion. In other words, there seem to be certain inherent soil properties that resist the erosive action of rainfall. An analysis of the physical characteristics which might influence the erosiveness of any particular soil suggests at least two ways in which erosion is affected. In the first place, the rate

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‡Superscript numerals refer to Bibliography, pages 44 and 45.

of water absorption and permeability of the soil profile determine to a great extent the amount of run-off for a given rain. Secondly, certain physical characteristics of the soil undoubtedly affect the amount of erosion taking place during run-off. It is necessary, therefore, to critically analyze the properties of the soil that affect the rate of water absorption, permeability of the profile, and the resistance of the soil particles to dispersion in running water.

The amount of water absorbed by the soil is influenced by its mechanical composition and structure. In sands and other light-textured soils, the coarseness of the particles is probably the most important factor contributing to a rapid water absorption and permeability and decreased erosiveness. In heavy soils, however, there apparently is no direct relationship between the clay content and the amount of erosion. For example, Bennett and Allison¹³ describe the Mantanzas clay from Cuba containing 90 per cent clay as well drained and friable; the Alto Cedra clay with only 52 per cent clay is pictured as poorly drained, having a stiff subsoil with a poor structure. The Mantanzas is distinctly granular and non-erosive; Alto Cedra exhibits no granular characteristics and is erosive. The degree of aggregation and the size and stability of the aggregates, therefore, must have a pronounced effect upon the rate of percolation and, consequently, upon the amount of run-off. The quantity of water absorbed by and percolated through the soil will undoubtedly be also affected by the amount of organic matter present and the swelling of the soil colloids.

Soils differ in the amount of material eroded per unit of run-off water. Carefully controlled experiments³² have shown that with rainfall, vegetation, and slope and area of the land constant, the amount of erosion per cubic foot of run-off decreased from .881 pounds to .477 pounds as a result of plowing under organic matter. These results suggest that the state of aggregation, and the size and stability of the granules play exceedingly important parts in decreasing the tendency of the soil to be dispersed by the force of running water. The ease of dispersion should be a paramount factor in soil erosion.

Most of our present knowledge concerning the factors contributing to soil erosion and erosion control has been based upon the results of field experiments. Some very valuable data on erosion control under different systems of management have been obtained^{5 18 32}. The use of erosion-preventing crops, terraces, soil-saving dams, strip farming and other practices has produced worth

while information. However, our knowledge of the physical and chemical properties of the soil which influence erosion has not been materially increased by these field experiments.

It is the object of this paper (1) to present the results of a laboratory study of the physical and chemical properties of several soils that are known to vary in their erosiveness, and (2) to show that the nature of the soil plays a most significant role in determining its susceptibility to erosion.

The data are presented in two parts: first, the physical properties of the soil that influence erosion, and, second, the physical and chemical properties of the extracted colloids and their relation to the characteristics of the soil.

PART I. PHYSICAL PROPERTIES OF SOILS AND SOIL EROSION

It should be expected that the extent of erosion would depend upon the amount and the velocity of run-off water since the carrying capacity of water increases as the sixth power of the velocity. The amount of run-off from any given rain depends upon the rate of water absorption and the permeability of the soil. The permeability of a soil profile is determined by the rate of percolation through the least permeable layer. In most upland soils the B horizon limits the permeability of the profile. Janota¹⁹ describes the B horizon of the podsol as the most impermeable layer since it contains the highest percentage of clay and has the lowest porosity. He found that soils with a total pore volume of less than 40 per cent were impermeable. Slater and Byers²⁴ studied the rate of percolation of water through a number of soil cores that were collected in such a way as to disturb the natural structure as little as possible. The rate of percolation was correlated with the silt content of the soil. No relationship existed between the order of permeability and the erosion indices as determined by the Middleton method²⁰. Baver⁸ observed that the percentage of non-capillary pores largely determined the drainage properties of soils. A soil with 5.9 per cent of air pores at a depth of 16 to 30 inches was well drained, but one with an air capacity of 2 per cent at the same depth was rather poorly drained. Kozeny²⁴ has expressed the rate of percolation through a soil as a function of the pore space, the viscosity of the medium and the dispersion of the soil. Antipov-Karatajev⁴ has reported that the filtration velocity through soils, in cubic centimeters of filtrate per minute, depends upon the nature

of the exchangeable ions present; the rate of filtration followed the series: $\text{Fe} > \text{Ca} > \text{Mg} > \text{NH}_4 > \text{Na}$. He also found that the rate of filtration was a function of the colloid content of the soil. Powell³³ observed that the oxide, sulphate, and carbonate of calcium had about the same beneficial effect on the permeability of the Putnam silt loam subsoil.

Middleton³⁰ studied the physical and chemical properties of three erosive and three non-erosive soils. He found that the properties giving the best index to erosiveness are the dispersion ratio, the ratio of colloid content to moisture equivalent, the erosion ratio, and the silica-sesquioxide ratio. The dispersion ratio was the most valuable single criterion. It is determined by shaking, end over end 20 times, a 10-gram sample of air-dry soil in a cylinder containing a liter of water. The dispersion ratio is the amount of silt and clay in suspension at a depth of 30 centimeters, expressed in percentage of the total silt and clay as determined by mechanical analysis. The erosion ratio is the quotient obtained by dividing the dispersion ratio by the ratio of colloid content to moisture equivalent. He classified soils with an erosion ratio of less than 10 as non-erosive. It is difficult to understand why any relationship should exist between soil erosiveness and such an arbitrary ratio, since the moisture equivalent is influenced by so many variables.

Several investigators^{1, 30} have shown that soils with a low silica-sesquioxide ratio are difficult to disperse. Bennett¹¹ observed a direct correlation between this ratio and the erosiveness of tropical soils. The ease of dispersion is also affected by the nature of the flocculating electrolytes present. Stevens and Cobb³⁵ leached soils with HCl and the carbonates and chlorides of Ca, Mg, Na, and K for three months and then determined their effect on the aggregation of the soil by means of an aggregate analysis⁷. They calculated the dispersion ratio from the data obtained and found that the ratio was lowered by the divalent cations and increased by the monovalent, including the H-ion.

Organic matter undoubtedly plays an important part in affecting the erosiveness of different soils. Bayer and Rhoades⁷ have shown that soils high in organic matter contain from 15 to 30 per cent more granules than those low in organic matter, and that the granules are three times more stable when shaken with water for 100 minutes. Jenny²² found that cultivation for forty years, as compared with a virgin prairie, caused a 38 per cent decrease in

organic matter, and a 28 per cent decrease in sand-size granules. The data of Jenny²⁰ and Bayer⁹ show a relative decrease in organic matter and granule content of soils from Canada to Texas, the percentage granulation of the silt and clay decreasing from 75 per cent in Canada to 25 per cent in Texas. The degree of saturation with calcium of these soils is practically the same. Burr and Russell¹⁷ report that the addition of organic matter increased the porosity of the soil and the stability of the granules. Data given by Middleton et al³¹ indicate an inverse relation between the organic matter content and the erosion and dispersion ratios. The organic matter apparently serves as a cementing material. The summation of all the effects of organic matter should lead to a decrease in soil erosion. These fragmentary data and the general observation that some soils erode more readily than others, indicate that the physical properties of soils have a great influence on erosion. However, the data have not been comprehensive enough to account for the observed differences in erosiveness of the soils studied.

Experimental Procedure

The Iredell sandy clay loam and the Davidson clay were selected for studying the physical properties of soils that should be expected to affect erosion. Field observations have definitely pointed out that the Iredell is an erosive and the Davidson a non-erosive soil. These two soils are characteristic of the Piedmont plateau region of North Carolina. They have usually been considered as weathering from the basic crystalline rocks without regard to any particular kind of rock. The two samples used in this work were selected from typical profiles of each type. An analysis of the underlying rock showed that the Iredell was weathered from diorite and the Davidson from gabbro. These soils occur in adjacent areas. Therefore, about all of the external factors which might affect erosion are practically constant. The marked differences in their erosiveness must be due to variations in their physical or chemical properties.

In order to determine the relationship between erosiveness and the soil properties which might influence erosion, the following physical properties were investigated: (a) the mechanical analysis by the pipette method, (b) aggregate analysis by a modification of the Kopecky elutriator method⁷, (c) moisture equivalent, and (d) dispersion and erosion ratios as calculated from the aggregate analyses data and as determined by the method of Middleton³⁰.

Mechanical Analysis and State of Aggregation of Iredell and Davidson Soils

It is interesting to note that there is a higher percentage of clay in all horizons of the Davidson than in the corresponding horizons of the Iredell. This is especially true in the A horizon. This horizon of the Iredell contains more than twice as much sand and 1.5 times more sand plus silt than the same horizon of the Davidson. Such a difference in texture might suggest that the Iredell was less erosive than the Davidson, especially since the B horizon of the Davidson contains more clay than the B layer of the Iredell. There is, however, a marked difference in the physical condition of the clay. The Iredell B horizon is a brownish-yellow, heavy, plastic, sticky and impervious clay when wet, which shrinks, cracks, and becomes very hard when dry. The Davidson, on the other hand, is a dark-red, heavy but friable clay which breaks into large lumps that are easily crumbled into small fragments or granules.

The difference in the character of the B horizon of these two soils is shown by the extent to which they are aggregated. Granulation was measured by the Kopecky type of elutriator⁷. The soils were dried just enough so they could be gently rubbed through a 2 mm. sieve, and were then placed in the elutriator without further treatment. In order to obtain ultimate dispersion they were shaken over night in a reciprocal shaker after treatment with a dispersing agent. The Iredell soil was dispersed with $\text{Na}_2\text{C}_2\text{O}_4$. The Davidson, however, flocculated in the presence of $\text{Na}_2\text{C}_2\text{O}_4$; consequently, NaOH was used as the peptizing agent. Although NaOH effectively dispersed the natural Davidson soil the electro-dialyzed colloid was not dispersed by any concentration of base. Apparently, the small amount of Cl -ions in the natural soil aided in obtaining dispersion. A considerable quantity of Cl -ions was found in the anode chamber during electro-dialysis.

A large part of the untreated Iredell soil was dispersed and carried over in the first few liters of water. Further elutriation with 17 liters of water did not carry over many fine particles. Only a small amount of the Davidson soil was carried over with the first few liters of water but continued elutriation caused a slow dispersion of the coarser aggregates. The material in the third tube was still cloudy after using 17 liters of water. Elutriation was continued until 34 liters of water had passed through the different tubes with the result that the third tube remained clear.

These observations indicate that there is practically no dispersed clay in the Davidson soil. The smaller particles that finally passed over represent the breaking down of rather instable granules. Consequently, the remaining aggregates include only those secondary units which are stable in an excess of water.

TABLE 1.—MECHANICAL COMPOSITION OF THE IREDELL AND DAVIDSON SOILS

Diameter of Particles mm.	Iredell Sandy Clay Loam			Davidson Clay		
	A	B	C	A	B	C
2-1	6.82	.47	.62	1.43	3.25	.47
1-5	6.52	.86	2.43	2.56	2.44	1.68
.5-.25	10.43	3.04	6.95	4.77	3.45	2.77
.25-.10	22.17	8.04	16.34	12.24	8.97	5.51
.10-.05	9.36	6.24	8.97	5.29	4.04	3.78
.05-.005	19.30	25.81	26.29	26.32	12.53	28.21
∠.005	24.40	56.09	38.40	47.42	65.39	56.60
∠.002	20.22	47.47	32.64	39.48	63.72	49.23

The data in Tables 2 and 3 clearly show that the Davidson is more highly granulated than the Iredell. This confirms Middleton's work³⁰ in which the dispersion ratio was found to be about 1.5 and 2 times higher in the Iredell A and B horizons, respectively, than in the corresponding horizons of the Davidson. The erosion ratio was two times greater in the A, and four times greater in the B horizon of the Iredell than in the same horizons of the Davidson.

TABLE 2.—AGGREGATE AND ULTIMATELY DISPERSED ANALYSES OF THE IREDELL SANDY CLAY LOAM

Size of particles (mm) and condition of aggregation	A horizon		B horizon		C horizon	
	Aggregate	Dispersed	Aggregate	Dispersed	Aggregate	Dispersed
2-1	13.48	8.97	5.34	.90	3.27	1.33
1-5	15.11	7.51	12.16	1.11	12.85	5.21
.5-.25	14.78	10.00	17.37	2.74	17.09	8.89
.25-.10	18.97	14.25	15.57	4.45	10.84	5.85
.10-.05	20.45	8.62	27.18	4.14	28.46	9.80
.05-.02	7.19	14.67	15.43	13.80	19.79	19.58
∠.02	10.03	35.98	6.97	72.86	7.72	49.35
∠.05	87.79	49.35	77.67	13.34	77.51	31.08
∠.25	43.37	26.48	34.87	4.75	33.21	15.43
Per cent of silt + clay in aggregates ∠.05 mm.-----	66.0		74.2		60.0	
Per cent of silt + clay in aggregates ∠ 25 mm.-----	33.3		34.9		25.8	
Per cent of total soil in aggregates ∠ 25 mm.-----	16.9		30.1		17.8	
Per cent of total soil in aggregates ∠.05 mm.-----	33.4		64.3		41.4	
Per cent of total aggregates present (∠.05 mm.) that are ∠ 25 mm.-----	50.5		46.9		47.9	

Figures 1 and 2 illustrate the degree of granulation of the two soils. A careful analysis of these graphs shows several very interesting and important facts concerning the extent and character of

TABLE 3.—AGGREGATE AND ULTIMATELY DISPERSED ANALYSES OF THE DAVIDSON CLAY

Size of particles (mm) and condition of aggregation	A horizon		B horizon		C horizon	
	Aggregate	Dispersed	Aggregate	Dispersed	Aggregate	Dispersed
2-1	10.30	1.78	28.43	2.47	25.10	.59
1-.5	20.81	3.88	32.63	2.73	22.33	2.31
.5-.25	21.80	6.30	17.17	3.88	14.59	2.91
.25-.10	14.18	8.14	7.63	6.50	7.48	2.43
.10-.05	20.76	8.67	7.70	4.88	14.75	2.45
.05-.02	6.09	17.36	1.79	2.59	6.83	5.46
∠.02	6.08	53.86	4.95	76.96	8.94	83.67
∠.05	87.85	28.77	93.56	20.46	84.25	10.69
∠.25	52.91	11.96	78.23	9.08	62.02	5.81
Per cent of silt + clay in aggregates ∠.05 mm.-----	82.9		91.9		82.5	
Per cent of silt + clay in aggregates ∠.25 mm.-----	57.5		87.0		63.0	
Per cent of total soil in aggregates ∠.25 mm.-----	41.0		69.2		56.2	
Per cent of total soil in aggregates ∠.05 mm.-----	59.1		73.1		73.6	
Per cent of total aggregates present (∠.05 mm.) that are ∠.25 mm.-----	69.2		94.6		52.9	

the aggregation of the A and B horizons of the profiles. If the total percentage of aggregates in the soil, as represented by the cross-hatched portion of the graphs, is considered, it is noted that both soils are rather highly granulated. Since the curves for ultimate dispersion and aggregate analysis cross at about .05 mm., this value has been chosen as the lower limit of aggregate size in calculating the total degree of aggregation. The equation for calculating the percentage of silt plus clay which is aggregated is:

$$\frac{\left\{ \sum \text{of particles } > .05 \text{ mm} \right\} \text{ without dispersion} - \left\{ \sum \text{of particles } > .05 \text{ mm} \right\} \text{ with dispersion}}{100 - \left\{ \sum \text{of particles } > .05 \text{ mm} \right\} \text{ with dispersion}} \times 100$$

= the per cent of silt + clay which is aggregated

The data in Tables 2 and 3 show that the Davidson contains 15 to 20 per cent more aggregates in each horizon than the Iredell. Nevertheless, there is a relatively high content of total aggregates in the Iredell. The B horizon* contains 64 per cent of granules, which is only 9 per cent less than the Davidson. It is seen that 91.9 per cent of the silt and clay in the Davidson is aggregated; in the Iredell only 74.2 per cent is in the form of granules larger in size than silt.

Neither the percentage of silt and clay aggregated, nor the percentage of total aggregates in the soil indicates the real differences between the two soils. If the B horizons are considered, it

* Since the B horizon limits the permeability of the soil profile all discussion of aggregation will be limited to this layer except where otherwise indicated.

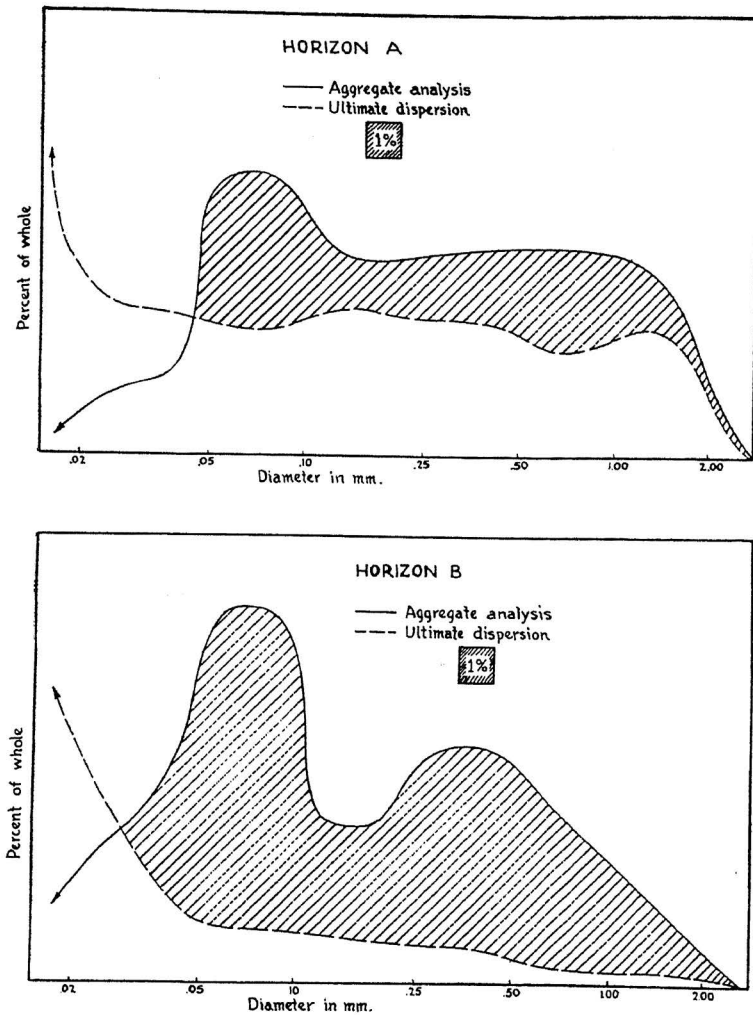


Fig. 1.—Size-frequency distribution curve of Iredell sandy clay loam with and without dispersion.

is observed that only 34.9 per cent of the silt and clay in the Iredell is aggregated into aggregates larger than .25 mm., whereas, in the Davidson 87 per cent of the corresponding mechanical separates is in the form of large granules (> .25 mm. in diameter). In other words, 149 per cent more of the silt and clay in the Davidson has been aggregated into large stable granules. Of the total aggregates present (> .05 mm.) 94.6 per cent are larger than .25 mm. in

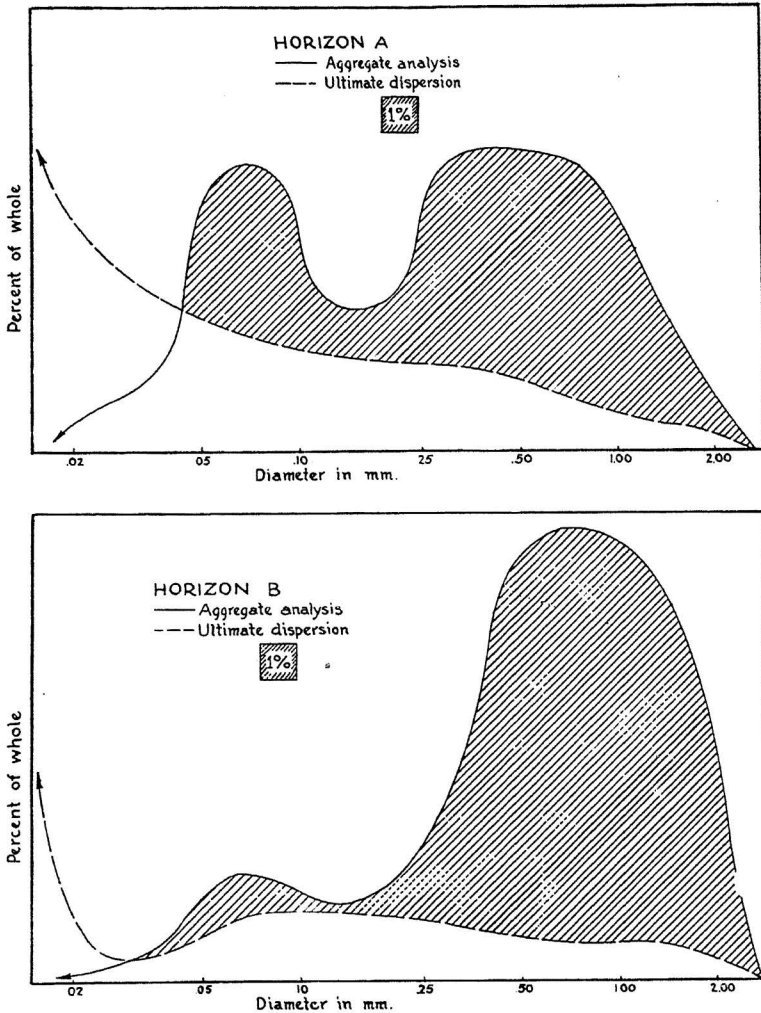


Fig. 2.—Size-frequency distribution curve of the Davidson clay with and without dispersion.

diameter in the Davidson; in the Iredell only 46.9 per cent are of the larger size. An analysis of the entire profile reveals that there is a more or less constant amount of aggregates in the Iredell from the surface downward. The Davidson, however, contains a higher percentage of large granules in the B horizon than in the A and C layers.

An examination of the granules with a wide field, binocular microscope shows that the character of the aggregates in these soils is markedly different. The Iredell aggregates appear to con-

sist of a sand grain which is coated with a layer of silt and clay. These sand, silt, and clay particles which make up the secondary particles are held together by some cementing material, probably gelatinous silica. This same material cements the aggregates together forming an impermeable soil layer.* When pressure is applied with the fingers to a cluster of these aggregates, they mold like putty except for the coarse, inner particle. They might be classed as "stone-fruit-like" aggregates. However, in the light of a recent classification of soil structure by Baver⁹ these secondary particles are not true granules and probably should be called fragments.

The Davidson aggregates are composed of groups of smaller aggregates that crumble under pressure into smaller granules. They appear distinctly porous under the microscope. They are true granules and may be referred to as the "pop-corn ball" type of aggregate.

Dispersion and Erosion Ratios

The dispersion ratio by the Middleton method³⁰ agrees somewhat with the aggregate analysis data. This should be expected since both methods determine to what extent the soil is dispersed by water. The values for the A, B, and C horizons of the Iredell are 4.6, 4.7 and 9.5 times higher, respectively, than those for the corresponding horizons of the Davidson (Table 4). The differences are greater than in the data reported by Middleton.

TABLE 4.—PHYSICAL PROPERTIES OF THE IREDELL AND DAVIDSON SOILS

Soil Type and Horizon	Colloid: mois- ture equiva- lent ratio	Mois- ture equiva- lent ratio	% Silt + clay dispersed	% Silt + clay not dispersed	Middleton Method		Aggregate Method*		Specific gravity
					Dis- persion ratio	Erosion ratio	Dis- persion ratio	Erosion ratio	
Iredell A	1.17	17.32	43.70	7.22	16.5	14.1	34.0	29.1	2.82
Iredell B	1.28	37.18	81.90	9.91	12.1	9.5	25.8	20.2	2.66
Iredell C	1.06	30.91	64.69	22.11	34.2	32.2	40.0	27.8	2.77
Davidson A	1.59	24.86	73.74	2.69	3.6	2.3	17.1	10.8	2.58
Davidson B	2.12	30.01	77.92	2.06	2.6	1.2	8.5	4.0	2.70
Davidson C	1.23	39.81	85.81	3.07	3.6	2.9	17.7	14.4	2.59

*In this method the dispersion ratio is = % silt + clay separated by aggregate analysis divided by total % silt + clay $\times 100$.

The erosion ratio is this dispersion ratio divided by the colloid:moisture equivalent ratio.

The erosion ratio suggests a greater difference between these two soils than any of the other determinations. The values of the erosion ratio (Table 4) for the A, B, and C horizons of the Ire-

*The swelling of the colloids undoubtedly plays an important role in the permeability of this layer. This will be discussed later.

dell are 6.1, 7.9 and 11.1 times greater, respectively, than for the same horizons of the Davidson. Middleton reports considerably lower values for the same soils with the exception of the C horizon of the Iredell. The erosion ratio, however, is a more or less empirical factor since it is difficult to visualize why any relationship should exist between the colloid: moisture equivalent ratio and the erosiveness of a soil.

The dispersion ratio should give a better index of the erosiveness of a soil because it is a measure of the ease of dispersion of the soil. Consequently, it will receive the greater emphasis in this discussion. The dispersion ratio as determined by Middleton is a measure of the stability of soil aggregates under the influence of moving water. Since an aggregate analysis measures essentially the same effects, the data thus obtained were used to calculate this ratio. The values were always higher than by the Middleton method. In other words, the aggregate analysis caused a greater dispersion of the secondary particles which may be due to the washing out of soluble flocculating electrolytes that are always present when using the technique of Middleton. Moreover, some silt and clay is probably flocculated orthokinetically when using sedimentation methods; in an aggregate analysis such a possibility is eliminated. Consequently, since these particular soils settle out almost entirely when using sedimentation methods of mechanical analysis one might expect the aggregate analysis method to give a better index of the real dispersion or aggregation of the soil.

Discussion

The aforementioned data show very clearly that one of the principal differences between erosive and non-erosive soils is the degree of aggregation of the finer mechanical separates into large, stable granules. The erosive soil (Iredell) contained 64.3 per cent of total aggregates ($>.05$ mm.) and the non-erosive (Davidson) 73.1 per cent. This difference is not great enough, however, to account for the observed variations in erosion. The data point out, however, that 69.2 per cent of the non-erosive and 30.1 per cent of the erosive soil is composed of granules larger than .25 mm. in diameter. Of the total aggregates present ($>.05$ mm.) 94.6 per cent and 46.9 per cent are larger than .25 mm. in the non-erosive and erosive soils, respectively. In the Davidson soil 87 per cent of the silt and clay is present in the form of granules $>.25$ mm.; in the Iredell soil only 34.9 per cent of the finer mechanical separates are aggregated into large granules. The Davidson aggregates are friable and porous; those in the Iredell are plastic and compact.

The high content of large porous aggregates in the Davidson is in marked contrast to the small, compact granules and dispersed condition of the Iredell. The size of the particles, whether primary or secondary units, determines to a great extent the amount of erosion by any given quantity of run-off.

The rate of percolation of water through the soil profile is one of the important factors influencing soil erosion. The physical nature of the B horizon determines largely the amount and rate of percolation since this is usually the limiting layer in permeability. A soil such as the Davidson with a high content of large, stable and porous aggregates which do not swell* should be expected to allow rapid percolation. The opposite, however, would be expected of a soil like the Iredell which contains a large amount of easily dispersed clay and a small quantity of coarse aggregates in both the A and B horizons. Experimental observations have confirmed this hypothesis. The B horizon of the Davidson was found to be nine times more permeable than the same layer of the Iredell. The soils were sieved through a 2 mm. sieve, after they had been partially dried. This drying and sieving apparently made the Iredell more granular than in its natural state and, consequently, the rate of percolation was increased. Slater and Byers³⁴ found even greater differences in the permeability of these two soils when the A horizon was used. Their data show that undisturbed cores of the Davidson were 34 times more permeable than those of the Iredell. Packing the soils in tubes decreased the difference between the rate of percolation of these two soils indicating that the physical condition of the Iredell is naturally unfavorable for rapid percolation. They probably would have observed a much greater variation with the B horizons because the difference in the degree of aggregation is much greater in this layer and, since the Iredell contains a relatively higher percentage of clay. The naturally unfavorable structure of the Iredell results in an increased amount of run-off and erosion. The Davidson, however, in its natural state is granular, friable and porous, and because the granules are stable under the influence of water, it is very permeable, which results in a reduction of the amount of run-off and erosion. Therefore, because of (1) the high degree of aggregation, (2) the large size and porosity of the granules, and (3) the stability of the granules in the Davidson soil, it has a high

* See part II.

absorptive capacity and permeability for water and resists the dispersing action of run-off water. Consequently, it is a non-erosive soil as contrasted with the erosive Iredell.

The factors responsible for the difference in the aggregation of these soils have not been definitely isolated. Studies by Bennett¹¹ and Anderson and Mattson² indicate that there is a correlation between the silica-sesquioxide ratio of the colloids and certain physical properties of soils. For example, the total exchange capacity, the absorptive capacity of the colloid for water vapor, the volume increase of the soil in water, and the volume of the flocculated colloid decrease with this ratio. The activity of the colloidal particles apparently is diminished as the content of iron and aluminum is increased. Since these physical properties vary with the silica-sesquioxide ratio of the colloids, the erosiveness of soils may be either directly or indirectly affected by the relative amounts of iron and aluminum present. Bennett's field observations point out that soils with a low silica-sesquioxide ratio are non-erosive. He describes a Costa Rican soil—the Columbiana clay—as exceedingly fine in texture, but “as friable as the mellowest loam, and in the practical sense is not susceptible to erosion.” The silica-sesquioxide ratio of this soil is 0.65. The Davidson clay is similar in its properties to the Costa Rican clay; the lateritic properties of the Davidson, however, are not as well defined as the soils described by Bennett.

Lateritic soils probably contain free iron and alumina which at the existing pH are precipitated as hydrated iron and alumina gels, a larger part of which eventually becomes dehydrated. That part which is not thoroughly dehydrated may serve as cementing material in the formation of large granules. Microscopical observations of the secondary units from the Davidson indicate such a possibility. Several investigators^{2 3 27} have shown that the charge of the particles decreases with increasing amounts of iron and aluminum. Data in Part II of this paper show that these lateritic types of colloids are not hydrated. Since hydration and charge are considered as the primary factors responsible for the stability of colloidal suspension²⁵ the Davidson colloids should be flocculated rather than dispersed.

Since the Iredell weathers from minerals high in sodium and potassium and the Davidson from minerals low in these bases but high in calcium and magnesium, it would be expected that the divalent bases in the Davidson would produce large stable aggregates,

whereas the sodium and potassium in the Iredell would cause dispersion. The percentage saturation with total bases is about the same in both soils. These effects are being further studied.

In order to obtain more information concerning the factors that influence granulation and permeability in the Iredell and Davidson soils, studies of the properties of the colloids extracted from the B horizons of these soils have been made. The results of these investigations will be discussed in Part II.

PART II. PHYSICO-CHEMICAL STUDIES OF THE EXTRACTED COLLOIDS

Introduction and Experimental Procedure

Most of the physical and chemical properties of soils are dependent upon or associated with the colloidal fraction. Therefore, much valuable information concerning the erosive properties of soils should be obtained from studies of the extracted colloidal material. Little attempt has been made to interpret the results obtained in studying soil colloids in their relation to soil erosion. A part of the data presented by various investigators^{3 & 28} can be used to explain indirectly some of the observed variations in the erosiveness of different soils. Most of these findings, however, have been of such a nature that they can only be directly interpreted as affecting soil erosion after further studies, such as permeability and degree of aggregation, have been made of the soil as a whole. In this investigation the physical properties of soils affecting erosion (Part I) are interpreted in light of a series of data upon the physico-chemical properties of the extracted colloids when treated with various cations.

The various colloids were investigated from the standpoint of (1) their chemical composition, (2) their saturation capacity for bases, (3) the effect of different exchangeable cations on (a) their state of flocculation or dispersion as related to charge and hydration, (b) their relative degree of hydration as indicated by viscosity and swelling measurements, and (c) their permeability to water. The relationship of these data to each other and to the physical properties of the soils from which they were extracted permit for the first time an insight into the physical and chemical nature of the soil as it affects erosion.

In addition to the colloidal materials extracted from the B horizons of the Iredell and Davidson, Putnam clay and bentonite

were used in certain phases of these investigations. The latter colloids were used with the hope that the numerous data that have been accumulated in other studies with these systems would aid in explaining some of the results obtained with the Iredell and Davidson colloids. The undialyzed colloids were concentrated with a Pasteur-Chamberlain filter, dried, ground in an agate mortar and analyzed chemically. The results of these analyses are given in Table 5. The chemical composition of the bentonite was not determined. The electrodyalized clays were diluted with distilled water and completely dispersed by vigorous stirring. The saturation capacity of the colloids for bases was determined conductometrically and potentiometrically according to the Bradfield method¹⁶.

TABLE 5.—THE CHEMICAL COMPOSITION OF THE COLLOIDS EXTRACTED FROM THE IREDELL, DAVIDSON AND PUTNAM SOILS

Constituent	Iredell†			Davidson* ‡			Putnam B horizon†
	A	B	C	A	B	C	
SiO ₂ -----	38.12	40.81	39.72	33.87	36.44	33.80	50.22
Al ₂ O ₃ -----	26.65	27.83	27.34	32.77	30.19	29.56	27.65
Fe ₂ O ₃ -----	19.20	17.26	17.31	16.39	20.40	23.65	4.74
MnO-----	.16	.02	.01	.51	.28	.08	.011
CaO-----	.83	.71	.77	.50	.52	.34	.92
MgO-----	1.03	.99	.82	.93	.64	.56	1.52
K ₂ O-----	.17	.16	.17	.27	.05	.19	.85
Na ₂ O-----	.06	.03	.07	.18	.20	.08	.449
TiO ₂ -----	.85	.76	.69	.71	.64	1.01	-----
P ₂ O ₅ -----	.64	.58	.36	.27	.25	.45	-----
Volatile matter-----	12.64	12.27	12.80	15.40	13.02	12.85	13.37
SiO ₂ /Al ₂ O ₃ -----	2.43	2.49	2.46	1.76	2.05	1.95	3.08
SiO ₂ /R ₂ O ₃ -----	1.66	1.78	1.76	1.33	1.44	1.28	3.02
Saturation capacity of colloid-----	-----	35	-----	-----	12	-----	60

*It was necessary to use NaCH to obtain dispersion of the Davidson. Some of the original replaceable cations were probably replaced and removed upon concentration with the Pasteur-Chamberlain filter.

†Analysis by Bradfield (15). Soil from the same field as the one used in this study.

‡Credit is due Mr. Ralph W. Cummings, former student assistant in soils at North Carolina State College, for assisting with these analyses.

H-clay suspensions were saturated with Li-, Na-, K-, Ca, and Ba-ions by adding the respective hydroxides. The bases were added slowly with constant stirring. All sols were diluted to 2 per cent by weight, placed in pyrex flasks, stoppered with paraffined corks and allowed to stand for about a month to reach equilibrium. These sols were used for the viscosity flocculation, charge, and permeability measurements. Part of these sols were evaporated to dryness, ground to pass a 300-mesh sieve and used in the swelling determinations.

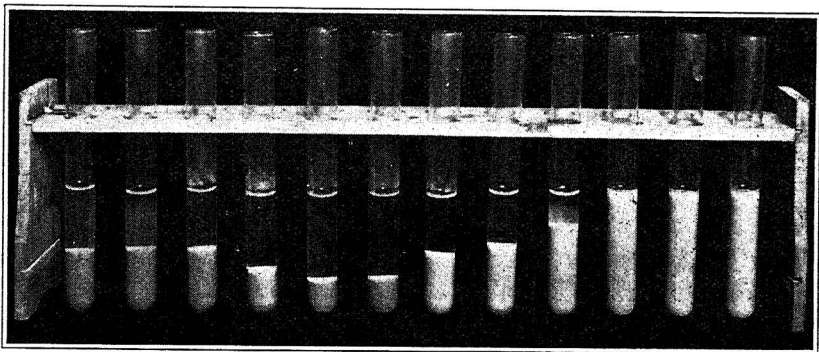
Chemical Studies of the Iredell, Davidson and Putnam Colloids

The data in Table 5 show that the Davidson is a lateritic soil if a SiO₂-Al₂O₃-ratio of 2 or less as suggested by Harrassowitz is

used as the criterion for lateritization. The Iredell with a ratio of 2.49 is intermediate between the Davidson and the Putnam, which has a $\text{SiO}_2\text{-Al}_2\text{O}_3$ ratio of 3.02. Apparently the impervious nature of the Iredell profile has not permitted as rapid weathering as the more permeable Davidson soil. The decrease in the $\text{SiO}_2\text{-Al}_2\text{O}_3$ ratio from the Putnam to the Davidson is accompanied by a decrease in the total cationic exchange capacity of the complex from 60 to 12 milliequivalents per 100 grams of colloid. This is in agreement with the data of several other investigators^{3 27}. The magnitude of the charge on the particles has also been shown to decrease with the silica-sesquioxide ratio^{2 3 27}. Since the intensity of the charge of the colloidal particles is considered as one of the main factors influencing the stability of colloidal suspensions²⁵ soils with a low silica-sesquioxide ratio such as the Davidson should be expected to flocculate. Soil with a ratio as high as the Putnam (3.02) should be stable.

Stability of Colloidal Clay Sols

It was observed throughout this study that all colloidal suspensions of the Davidson were flocculated irrespective of the nature of the exchangeable cations on the complex. Only the H-, Ca-, and Ba- suspensions of the Iredell and Putnam were flocculated; these suspensions precipitated very slowly as compared with the Davidson sols. The flocculation time (Table 6) was considered as that moment when a thin layer of supernatant liquid could be observed above the suspension. Figure 3 shows the volume of the floc of the different flocculated sols after standing 24 hours. The volume per gram of clay in the flocculated systems, the pH,



Li Na K H Ca Ba - H Ca Ba - H Ca Ba
 Davidson Sols Iredell Sols Putnam Sols
 Fig. 3.—The degree of flocculation of different clay sols after 24 hours.

and the zeta-potential and migration velocity of the colloidal particles are tabulated in Table 6.

TABLE 6.—THE RELATION OF DIFFERENT CATIONS TO THE REACTION, CHARGE, AND FLOCCULATION OF SEVERAL COLLOIDAL CLAYS

Colloid and Treatment	pH	Zeta potential millivolts*	Migration velocity microns per sec. 1 volt per cm.	Flocculation time	Volume of floc. after 24 hours cc./gm.
H-Iredell.....	4.04	39.0	3.07	4 min.	5.2
Li-Iredell.....	6.91	42.1	3.31	Stable	
Na-Iredell.....	7.29	41.2	3.24	Stable	
K-Iredell.....	6.91	38.9	3.06	Stable	
Ca-Iredell.....	6.59	35.8	2.82	5 min.	6.0
Ba-Iredell.....	6.58	29.7	2.34	6 hrs.	7.2
H-Davidson....	4.50	36.8	2.90	1 min.	4.0
Li-Davidson....	6.85	40.5	3.19	2 min.	5.2
Na-Davidson....	7.31	39.4	3.10	2 min.	5.4
K-Davidson....	6.92	37.9	2.98	2 min.	5.4
Ca-Davidson....	7.13	36.3	2.86	1 min.	3.0
Ba-Davidson....	7.11	35.3	2.78	1 min.	3.2
H-Putnam.....	4.10	36.3	2.86		
Li-Putnam.....	7.61	40.3	3.17		
Na-Putnam.....	7.66	38.5	3.03	All stable after 24 hrs.	
K-Putnam.....	7.32	38.7	3.05	H, Ba, and Ca flocculated in about 3 days	
Ca-Putnam.....	6.96	35.0	2.75		
Ba-Putnam.....	6.86	32.6	2.56		

*Credit is due Mr. R. F. Reitemeier for making the cataphoretic measurements.

An analysis of these data and associated factors throws much light on the relation of charge and hydration to the flocculation of colloidal clay suspensions. Kruyt²⁵ points out that the two factors influencing the stability of colloidal suspensions are the charge and the hydration of the dispersed particles. Wiegner³⁶ and Bradford¹⁴ consider the magnitude of the charge of clay particles as being due to the anionic nature (inner layer) of the colloidal complex and that the potential increases or decreases directly with the extent of dissociation of cations from the colloidal surface.

Mattson²⁷, Anderson³ and Anderson and Mattson² have shown that the charge and the exchange capacity of the colloid vary directly with the silica-sesquioxide ratio. Mattson²⁸, however, has shown that in the presence of NaOH colloidal particles charge themselves to a potential of 56 millivolts regardless of the exchange capacity.

The charge of the Iredell, Davidson and Putnam colloids was determined ultramicroscopically by means of a modification of the Tuorila cell⁶. Suspensions of 0.0001 per cent by weight were used; the results are given in Table 6. The zeta-potential was calculated by the Helmholtz-Perrin equation, using the factor 4. The results with the Putnam sols agree with those reported by Bayer⁶. The potentials follow the ionic series: Li > Na = K > H > Ca > Ba. The Li-, Na-, and K-Putnam sols were all stable but the H-, Ba-,

and Ca- systems flocculated very slowly, as shown in Figure 3. The intensity of the charge of the lateritic Davidson and the semi-lateritic Iredell colloids do not agree, however, with the data of Mattson²⁷, Anderson³, and Anderson and Mattson². From the data of other investigators^{27 3 2 36 14} these colloids, especially the Davidson, would naturally be expected to have a lower negative charge than the putnam. For example, Anderson³ reports that the Na-saturated colloid from the Marshall silt loam ($\text{SiO}_2/\text{Al}_2\text{O}_3=3.46$ and saturation capacity of 60 m. e./100 gms.) has a migration velocity of -4.8 microns per second; the Na-saturated Sassafras colloid ($\text{SiO}_2/\text{Al}_2\text{O}_3=2.35$ and saturation capacity of 24.3 m. e./100 gms.) has a migration velocity of -2.7 microns per second; the Na-Cecil colloid ($\text{SiO}_2/\text{Al}_2\text{O}_3=1.44$ and saturation capacity of 9.8 m. e./100 gms.) has a migration velocity of -2.4 microns per second. The $\text{SiO}_2\text{-Al}_2\text{O}_3$ ratios and exchange capacities of the Marshall, Sassafras and Cecil colloids compare very closely with those of the Putnam, Iredell and Davidson, respectively. The negative potentials of all Iredell and Davidson sols were higher than those of the Putnam with the exception of the Ba-Iredell and the K-Davidson. It is difficult to explain these striking differences in light of existing information. Although this problem is being further investigated certain facts have been observed in the present study which may be used to partially explain these puzzling results.

In the first place, the large differences in the hydration of the Davidson, Iredell and Putnam colloids may affect the relative migration velocities of the particles. The method of calculating the zeta-potential from the Helmholtz-Perrin equation does not take into consideration the hydration of the particles. The data in Table 7 show that the Putnam systems are more highly hydrated than those of the Iredell; the Davidson is not appreciably hydrated. It is possible that hydrated particles move somewhat slower in an electric field than particles which have no "water hull." In other words, the presence of a water shell around the particle might decrease the effectiveness of the "active points" on the surface of the colloid. Consequently, even though the Putnam colloid has five times as many "active points" (evidenced by its higher exchange capacity) as the Davidson, the high hydration of the Putnam might directly or indirectly impede the movement of the particles in an electric field.

Secondly, symmetry value* determinations show that the H-ions are more easily replaced from the Davidson and Iredell col-

* The symmetry value is the milliequivalents of H-ions replaced by K-ions at symmetry concentrations, expressed in percentage of the total milliequivalents of H-ions present. See Jenny (21).

loids than from the Putnam. In other words, the Putnam colloidal clay acid is less "dissociated" than the Iredell and Davidson clay acids.

The symmetry values are:

	Putnam	Iredell	Davidson
With same clay concentration (KCl concentration=exchangeable H-ions)	12.33%	18.85%	20.0 %
With same KCl concentration (same as exchangeable H-ions)	13.52%	16.66%	20.33%

The values for the Putnam colloid agree very closely with those reported by Jenny²¹ if allowance is made for slight differences in the clay concentrations. These results show that about 20 per cent of the replaceable H-ions on the Davidson colloid are exchanged by K-ions when the total concentration of both ions in the system is the same. Only about 13 per cent of the H-ions on the Putnam colloid are exchanged by K-ions at the symmetry concentration. In other words, the H-ions on the Davidson are more active than those on the Putnam when considered on the basis of the total quantity present. Determinations of the pH values of the electro dialyzed colloids also indicate that the clay acids in the Davidson and Iredell are more highly "dissociated" than the Putnam acid clay. Calculations based upon the pH values show that 1.26, 1.24, and 0.66 per cent of the total H-ions present are "dissociated" in the Davidson, Iredell and Putnam clays, respectively. If the "dissociation" of the Davidson is taken as 100, the Putnam and Iredell clay acids are 52.4 and 98.4 per cent "ionized", respectively. The symmetry values show that the exchangeable H-ions on the Putnam and Iredell colloids are only 61.4 and 94.4 per cent, respectively, as easily replaced as those on the Davidson. The two methods, therefore, give similar results and suggest that the potential of the Davidson colloid may be similar to the Putnam because of a greater activity of the exchangeable ions.

Thirdly, Lottermoser and Riedel²³ have shown that dilution of the positive Al_2O_3 , Fe_2O_3 , Cr_2O_3 and ThO_2 sols caused a reversal of charge; the sols were negatively charged in dilute concentrations. They attribute these results to an increase in dissociation with dilution and state that dilution and dialysis have the same effect on the intensity of the charge of colloidal particles. The sols used in the present investigation were rather dilute (0.0001%)

and, in light of the dilution effect reported by Lottermoser and Riedel, the Davidson and Iredell colloids may exhibit a higher negative charge than would be expected in more concentrated systems. The unexpected values of the zeta-potential may be due in part, therefore, to (1) the greater hydration of the Putnam colloids, (2) a higher activity of the exchangeable ions on the Iredell and Davidson colloids as shown by the symmetry value and pH measurements, and (3) the effect of dilution on the increase of the negative charge of colloidal particles as shown by Lottermoser and Riedel.

The flocculation of the Davidson sols irrespective of the amount and nature of cations present, and in spite of a high zeta-potential indicates that hydration is probably a more important factor influencing the stability of colloidal clay suspensions than is usually considered. The non-hydrated condition of lateritic sols such as the Davidson, allows them to maintain a high state of granulation which is not destroyed by the influence of water because of their inability to become hydrated. Semi-lateritic and non-lateritic colloids such as the Iredell and Putnam are capable of becoming rather highly hydrated and, therefore, probably form very unstable secondary particles except under extreme dehydration or in the presence of cementing materials such as organic matter. These results on the relative stability of the Iredell and Davidson colloids point out quite distinctly the reasons for the high degree of aggregation of the Davidson and the resistance of its granules to the dispersing action of run-off water.

Hydration of Colloidal Clays

It has been pointed out that the hydration of colloidal clay is an important factor influencing the stability of the colloidal system. Several methods of measuring the degree of hydration of soil colloids may be used depending on whether the colloid is in the dry state or in suspension. Hydration can be determined by measuring the liquid intake of dry colloids³⁷. The significance of this type of water will be discussed in connection with the swelling data. The relative hydration of different sols may be indirectly determined by calculations from viscosity data.

Viscosity of Clay Suspensions.—Viscosity determinations were made with the 2 per cent suspensions of Iredell, Davidson, Putnam, and bentonite colloids. In addition, 2.3534 per cent suspensions were used in order that comparisons might be made with the data reported by Bayer⁶ on suspensions of Putnam clay of the same

concentration. A Washburn modification of the Ostwald type of viscosimeter was used. Since the Davidson sols were all flocculated considerable difficulty was experienced in obtaining reliable results. Trial experiments showed, however, that duplicate results could be obtained with the use of regulated suction. The applied suction changed the time of flow of water through the viscosimeter from 193.8 ± 1.2 to $69.5 \pm .25$ seconds. All measurements were made in a thermostat at 30°C . Values for the relative viscosity and the grams of water associated with a gram of clay are given in Table 7.

The water associated with the clay, as calculated from viscosity measurements, may consist either of a water "hull" around the clay particles (if completely dispersed) or of occluded water that is mechanically held between the particles in a flocculated system. The thickness of the water hull undoubtedly is related to (1) the adsorptive capacity of the colloid for water molecules as affected by the nature of the inner layer and the amount and kind of exchangeable cations in the outer layer of the colloid and (2) the concentration of the sol. The amount of occluded water depends largely upon the degree of flocculation of the system, but is probably affected also by the hydration of the flocculating cation. In systems such as the highly flocculated Davidson sols most of the water is probably held as occluded water. In thoroughly dispersed systems most of the hydration water is present as a shell around the colloidal particle. Calculations of the hydration water were made according to the Meyer and Mark modification²⁹ of the Einstein equation²⁵:

$$\frac{N_s}{N_o} = 1 + 2.5 \frac{\phi}{V - \phi} \text{ where,}$$

ϕ = the effective volume of the dispersed phase,

V = the total volume of sol,

$$\frac{N_s}{N_o} = \text{the relative viscosity of the sol.}$$

The data in Table 7 show that the various cations do not affect the viscosity of different clays in the same order except that the viscosity of the H-, Ba- and Ca- sols always ranked in the order named. The high content of water associated with a gram of clay in the Davidson systems and in the H-, Ca-, and Ba Iredell sols is due to water of occlusion since all of these sols were completely flocculated. The H-, Ba-, and Ca- Putnam sols flocculated so slowly that the amount of occluded water was probably very small. The data also show that the Li-, Na- and K-Putnam colloids are more highly hydrated than the corresponding Iredell systems. The

TABLE 7.—THE RELATION OF EXCHANGEABLE CATIONS TO THE VISCOSITY AND SWELLING OF SEVERAL COLLOIDAL CLAYS

Colloid and Treatment	Relative viscosity	Water of hydration* cc per gm.	Liquid intake (cc/gm.)		
			H ₂ O	C ₆ H ₆	Swelling
H-Bentonite	7	35	3.77	1.45	2.32
Li-Bentonite	4.41	28.41	13.45	1.35	12.10
Na-Bentonite	3.41	24.11	12.27	1.20	11.07
K-Bentonite	2.92	21.31	9.76	1.20	8.56
Ca-Bentonite	3.42	24.16	4.00	1.50	2.50
Ba-Bentonite	6.51	33.96	3.90	1.40	2.50
H-Putnam	1.21	3.49	1.80	.98	.82
Li-Putnam	1.23	3.84	6.51	1.25	5.26
Na-Putnam	1.25	4.19	5.45	1.25	4.20
K-Putnam	1.24	4.04	1.46	.96	.50
Ca-Putnam	1.19	3.19	1.90	1.00	.90
Ba-Putnam	1.20	3.34	1.74	.90	.84
H-Iredell	1.51	8.08	1.69	1.46	.23
Li-Iredell	1.17	2.83	1.78	1.36	.42
Na-Iredell	1.16	2.63	1.98	1.38	.60
K-Iredell	1.18	2.98	1.53	1.51	.02
Ca-Iredell	1.27	4.53	1.92	1.62	.30
Ba-Iredell	1.26	4.33	1.86	1.50	.36
H-Davidson	1.41	6.66	2.26	2.41	-15 [†]
Li-Davidson	1.58	9.01	2.02	2.18	-16
Na-Davidson	1.51	8.06	1.90	2.08	-18
K-Davidson	1.77	11.41	1.93	2.35	-42
Ca-Davidson	1.31	5.11	1.94	2.22	-28
Ba-Davidson	1.35	5.76	2.08	2.43	-35

*Calculated by the Mark and Meyer equation. (See context).

†The lower intake of water is due to a settling of the fluffy, dry colloid when wetted.

chemical nature of the colloidal particle is undoubtedly responsible for these differences. Apparently, the inner layer of the Putnam colloid has a relatively high polarizing effect on the water molecules. All bentonite sols were extremely viscous indicating that the chemical nature of the alumino silicate particle must have a pronounced effect upon the hydration of the particle. Bentonite has an exchange capacity of about 90 milliequivalents per 100 grams of colloid.

The effects of varying amounts of base on the viscosity of the Iredell, Putnam* and Davidson sols are shown in Figure 4. Small additions of NaOH to the H-Iredell caused a rapid decrease in the relative viscosity by dispersing the system and freeing occluded water. Further additions up to 125 per cent saturation caused no changes but above that point another decrease occurred. The constant value between the two breaks is probably due to the combined effects of freeing occluded water and increasing the water hull because of the highly hydrated Na-ions on the exchange complex. The decrease above 125 per cent saturation is due to a diminution in the water hull because of excess Na-ions. The Putnam and Davidson curves are similar, but are probably the result of

* Data reported by Bayer (6).

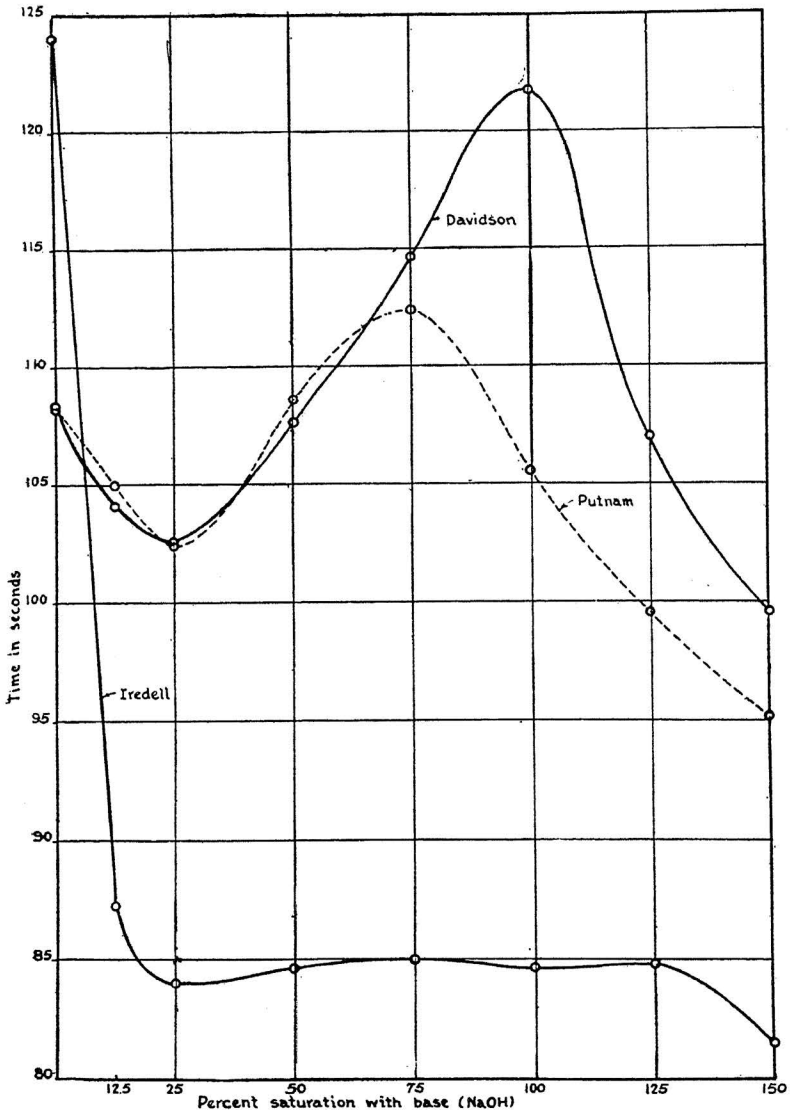


Fig. 4.—Viscosity of different clay sols with varying amounts of NaOH.

different causes. The first slight decrease is due to a dispersing effect, which in the Putnam is small because the H-Putnam is only slightly flocculated; in the Davidson it is small because the Nations did not cause complete dispersion. The rise in viscosity of the Putnam with further additions of base is caused by an increase

in the thickness of the water hull on the colloids; in the Davidson the rise is due to an increase in the water content (occluded water) of the floccules as a result of increasing quantities of hydrated Na-ions. An excess of NaOH causes a dehydration of the Na-ions on the complex which results in a decrease in the relative viscosity. These conclusions are based on differences observed in swelling and flocculation of the various systems, since it would be difficult to distinguish between occluded and hydration water from viscosity measurements alone. The viscosity determinations do indicate that the dispersed Putnam clay is about twice as highly hydrated as the dispersed Iredell clay.

Swelling of Soil Colloids.—Numerous investigators have noted a certain relationship between swelling and other properties of soil colloids. Bennett¹¹ observed that soils with a low silica-sesquioxide ratio were more friable, permeable, and resistant to erosion than those with a high ratio, and that the friable soils showed no visible swelling or shrinking even at extreme moisture variations. Soils with a high $\text{SiO}_2\text{-R}_2\text{O}_3$ ratio were subject to wide changes in volume during drying or wetting. Anderson³ found that swelling increased with the $\text{SiO}_2\text{-R}_2\text{O}_3$ ratio and that Na-saturated clay swelled about twice as much as K-clay. Additions of Ca-ions to the Na-clay decreased the amount of swelling. Mattson²⁸ studied the effect of adsorbed cations on the swelling of soils and found that swelling decreased according to the ionic series: Na > K > Ca > Mg > H > methylene blue. Winterkorn and Baver³⁷ have reported that the total amount of water taken up by the colloidal clays increased with the $\text{SiO}_2\text{-R}_2\text{O}_3$ ratio. They also show that water intake extends over a longer period of time in soils with a high $\text{SiO}_2\text{-R}_2\text{O}_3$ ratio.

The swelling data in Table 7 were obtained by the method of Winterkorn and Baver³⁷ in which the difference in the absorption of water and benzene is used as a measure of the swelling or hydration of the colloid. An analysis of the data show that the order of swelling of the different clays is: Bentonite* > Putnam* > Iredell > Davidson, with the latter exhibiting no swelling. The $\text{SiO}_2\text{-R}_2\text{O}_3$ ratios of these colloids decrease in the same order. Baver and Winterkorn¹⁰ have shown that the swelling of any given colloid varies with the nature of the cations on the exchange com-

* The swelling data on the bentonite and Putnam colloids were obtained by H. F. Winterkorn (37).

plex. In bentonite swelling followed the ionic series: $\text{Li} > \text{Na} > \text{K} > \text{Ca} = \text{Ba} > \text{H}$; in Putnam clay the series was: $\text{Li} > \text{Na} > \text{Ca} > \text{Ba} > \text{H} > \text{K}$; in the Iredell the series was $\text{Na} > \text{Li} > \text{Ba} > \text{Ca} > \text{H} > \text{K}$; the Davidson exhibited no swelling irrespective of the nature of the exchangeable cations present. The low swelling of the K-saturated Iredell colloids confirms the results of Baver and Winterkorn¹⁰ with other K-saturated clays. Moreover, their observations concerning the relative differences between the swelling of Ca- and H-saturated systems are confirmed by the behavior of the Ca- and H-Iredell colloids. A theoretical discussion of the effect of exchangeable ions on the swelling of soil colloids will be forthcoming in a paper by Baver and Winterkorn. The practical significance of these results will be mentioned in the general discussion of this paper. The data distinctly point out that the hydration of clays is to a great extent determined by the chemical nature of the clay particle and that colloidal clays behave differently in their swelling than bentonite.

The swelling data partly explain the differences between the rate of percolation of water through the Iredell and Davidson soils, and the extent to which the soils are aggregated into stable granules.

Inasmuch as the Davidson colloid does not swell in the presence of water one should expect the soil to have a relatively high permeability for water; the aggregates should be more stable than those secondary particles that increase in volume when wetted. Consequently, on the basis of hydration alone, the Davidson soil should be more resistant to the erosive action of water than the Iredell. The significance of hydration will be more vividly illustrated in the difference in the permeability of the Davidson and Iredell clay membranes.

Permeability of Colloidal Clay Membranes

The permeability of a soil profile is controlled largely by the amount and state of dispersity of the colloidal material present. Many soils with a highly colloidal B-horizon such as the Iredell and Putnam, are only slightly permeable to water; other soils, however, such as the Davidson with just as high a colloid content are extremely permeable. Bennett¹¹ observed that certain lateritic soils with a very high clay content were decidedly permeable to water and could be cultivated soon after heavy rains; these permeable clay soils apparently were not highly hydrated. Antipov-

Karatajev⁴ observed that the rate of percolation through a soil was a function of the colloid content and the nature of the exchangeable ions present. Soils saturated with iron were the most permeable. Baver⁵ studied the rate of ultrafiltration through Na- and Ca- colloidal clay membranes and calculated the size of the pores in the membranes on the basis of the rate of filtration. The pores in the Ca-clay membrane were found to be 8.3 times larger than those in the Na-clay.

In the present investigation the permeabilities of colloidal clay membranes of the Iredell, Davidson, Putnam and bentonite colloids, when treated with various cations, were studied. The clay membranes were formed by filtering 100 cc. of 1 per cent sols through collodion-filter paper membranes using an ultrafilter with a pressure of 100 pounds per square inch. The filters were so prepared as to permit the percolation of water at the rate of 2 to 3 cc. per minute. Trial experiments showed that the only effect of variations within this range was on the rate of filtration of the sol immediately at the beginning; no variations were observed after the clay membrane had been formed.

The rate at which the intermicellar liquid was removed and the rate of percolation of the ultrafiltrate through the clay membranes thus formed are shown in Figures 5 to 7, inclusive.

The data in these graphs show a segregation of the curves into two groups as far as the effects of the exchangeable cations are concerned. The one group contains the Li-, Na-, and K-sols and the other the H-, Ba-, and Ca-systems. This division is very pronounced with the Iredell and Putnam colloids; in the Davidson the distinction is not so great. Apparently, the rate of ultrafiltration is related to the degree of flocculation, the more flocculated systems being more permeable than the dispersed. Since all of the Davidson sols are flocculated, the rates of filtration are somewhat similar. Qualitative tests of the rate of filtration through a glass filter under a pressure of one atmosphere showed that all the Davidson systems and the H-, Ba-, and Ca-Iredell sols flocculated and that the volumes of the floccules formed remained practically constant until the supernatant liquid had been removed. The volume of the floc then decreased under the applied pressure resulting in the squeezing out of occluded water from the loosely-held floccules. The filtration up to this point apparently proceeded more or less as a water displacement through the flocculated mass. This displacement effect resulted in a constant rate of filtration until the

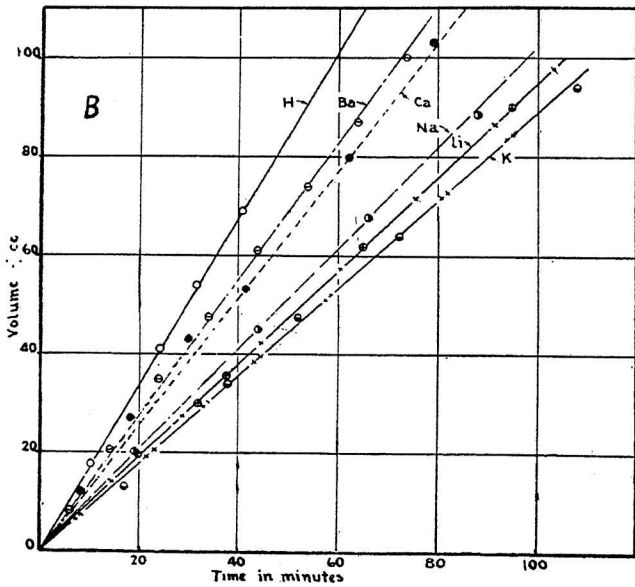
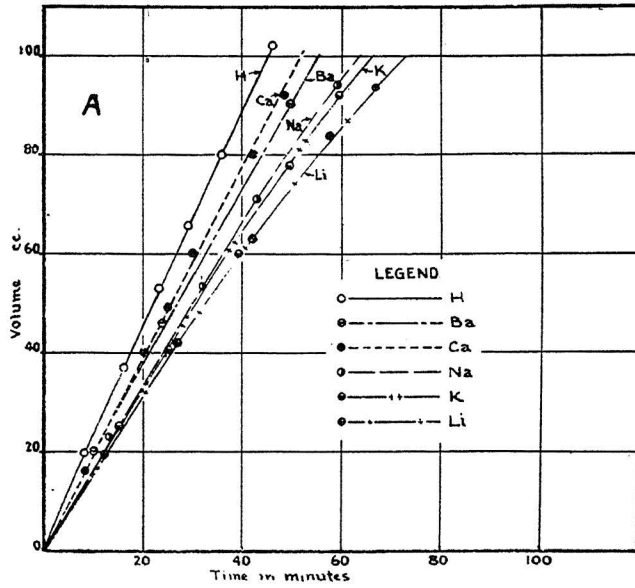


Fig. 5.—The effect of different cations upon the permeability of Davidson clay. (A) ultrafiltration of sol, (B) permeability of clay membrane.

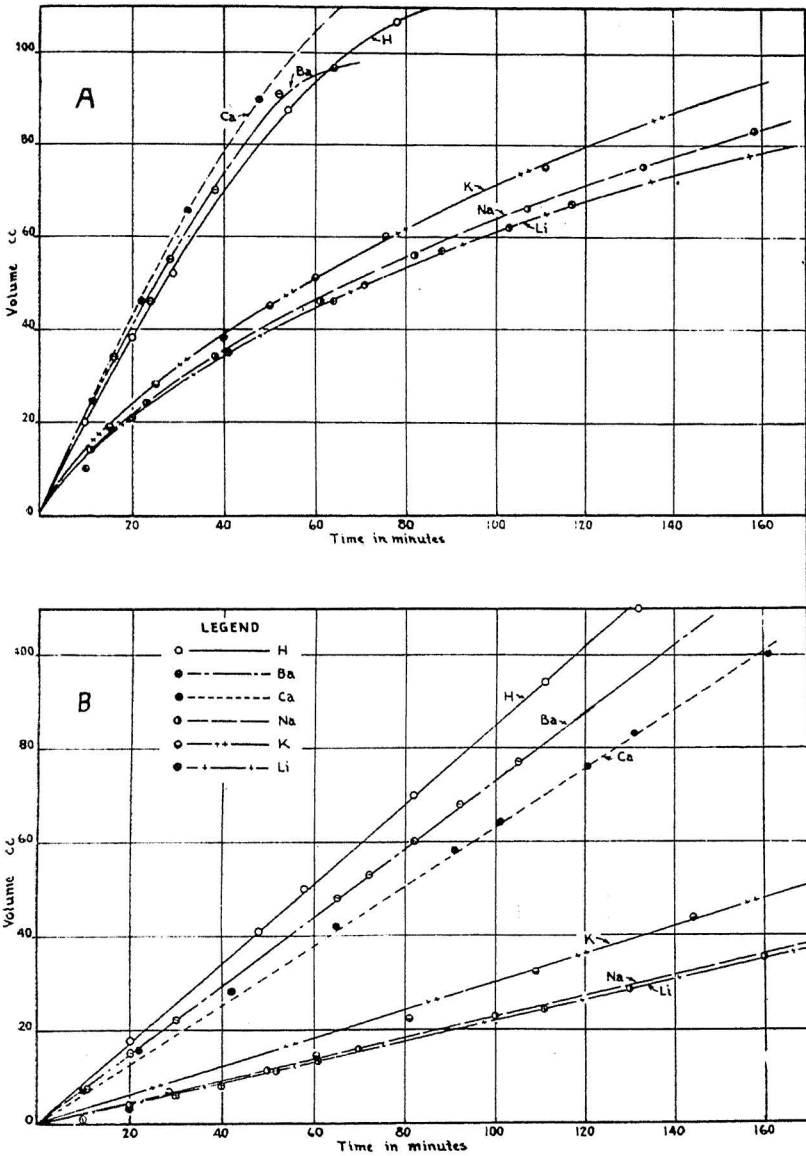


Fig. 6.—The effect of different cations upon the permeability of Iredell clay. (A) ultrafiltration of sol, (B) permeability of clay membrane.

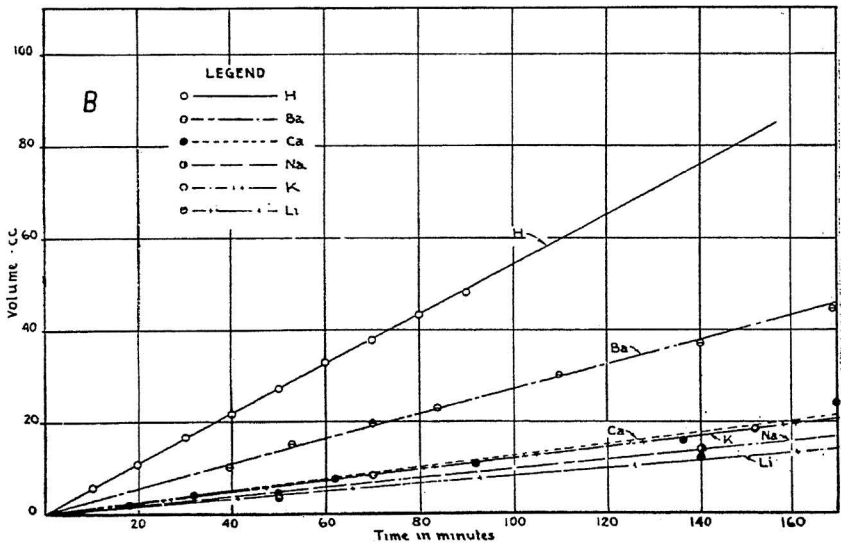
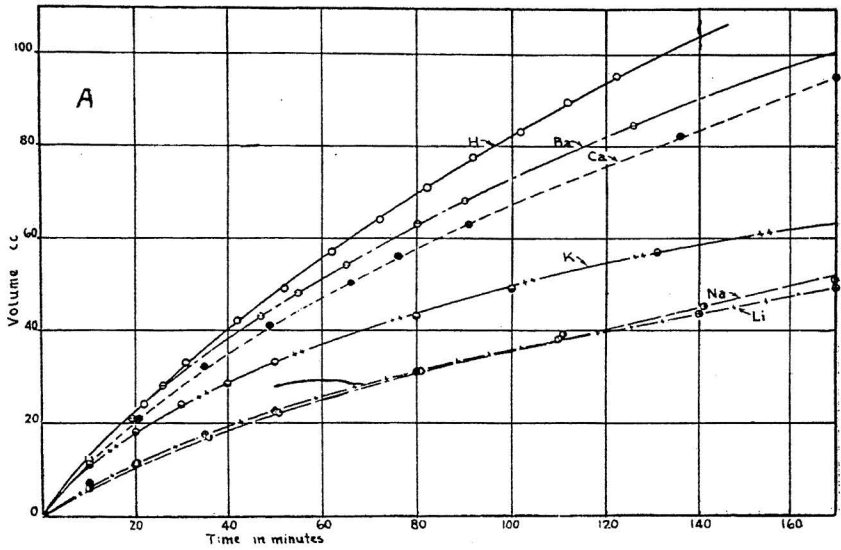


Fig. 7.—The effect of different cations upon the permeability of Putnam clay. (A) ultrafiltration of sol, (B) permeability of clay membrane.

last few cubic centimeters of intermicellar liquid were removed. Packing of the clay membrane by removal of occluded water then caused a decrease in the rate of percolation in the more hydrated systems such as the Ba- and H- Iredell; in the weakly hydrated systems such as the Davidson the rate of filtration remained constant until all of the intermicellar liquid had been removed.

The permeability of the membranes formed from one gram of clay or bentonite follows the definite order of $H > Ba > Ca$ which is the same as that of their relative viscosity. In the Iredell and Putnam colloids and in the bentonite, the order of the alkali cations was: $K > Na > Li$; in the Davidson the order was $Na > Li > K$. No definite relationship existed between the swelling and the permeability of the various clay membranes if all systems are considered. If, however, only the H-, Ca-, and Ba- systems are considered, permeability is inversely related to the amount of swelling. This is true with all colloids except the Iredell in which the differences in swelling of the various sols is vary small. Such a relationship should naturally be expected since swelling results in a decrease in the effective pore volume of the clay. H- clays were found to swell less than Ca- and Ba- clays; nevertheless, the H-systems were more permeable than those saturated with Ca- and Ba- ions. Apparently, the formation of granules that do not change their volume when wetted is an important factor in soil permeability. The fact that H-clays swell less and are more permeable than Ca-saturated clays throws considerable doubt upon the direct effect of liming acid soils on soil granulation and permeability.

It is noted that K-saturated clays occupy an intermediate position between the Na- and Li- systems and those saturated with H-, Ca-, and Ba- ions. The higher permeability of the K-clays over the Na- and Li-systems is due to the low swelling of the former. Since the K-colloids are dispersed their relative permeabilities are not as high as the flocculated H-, Ca-, and Ba- suspensions. The data showing the effect of the exchangeable cations on the permeability of clay membranes indicate that the rate of percolation is dependent upon the degree of dispersion and extent of swelling of the colloids.

Comparisons of the effect of the chemical nature of the colloidal complex on the permeability of clay membranes are shown in Figures 8 to 10 inclusive. These results show that irrespective of the nature of the adsorbed cation the order of permeability for the different colloids is: Davidson > Iredell > Putnam > bentonite.

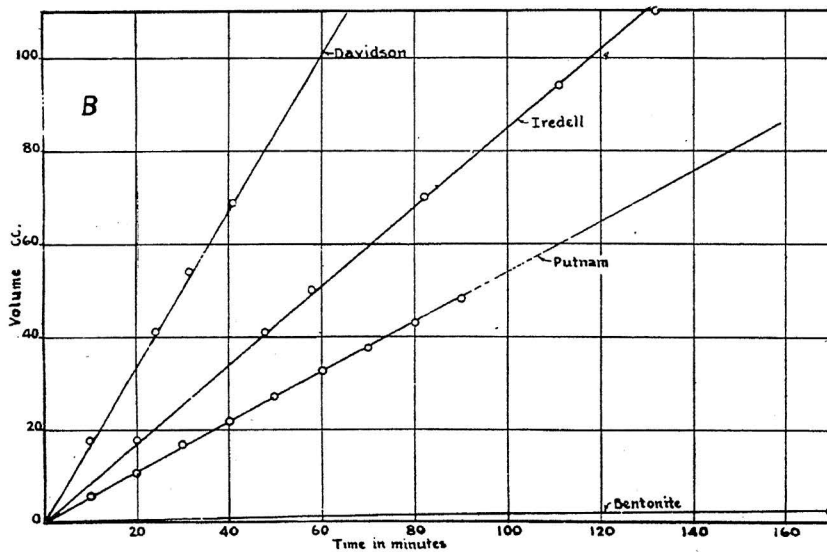
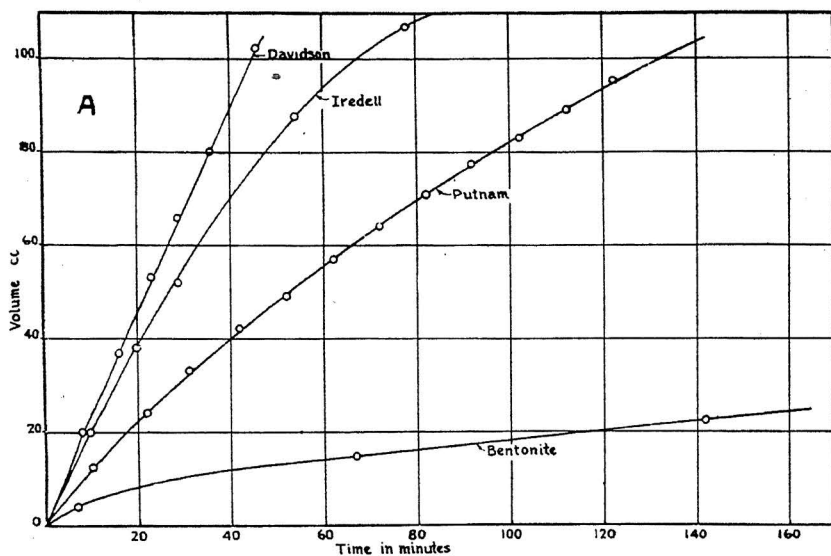


Fig. 8.—The permeability of different clays saturated with H-ions. (A) ultrafiltration of sol, (B) permeability of clay membrane.

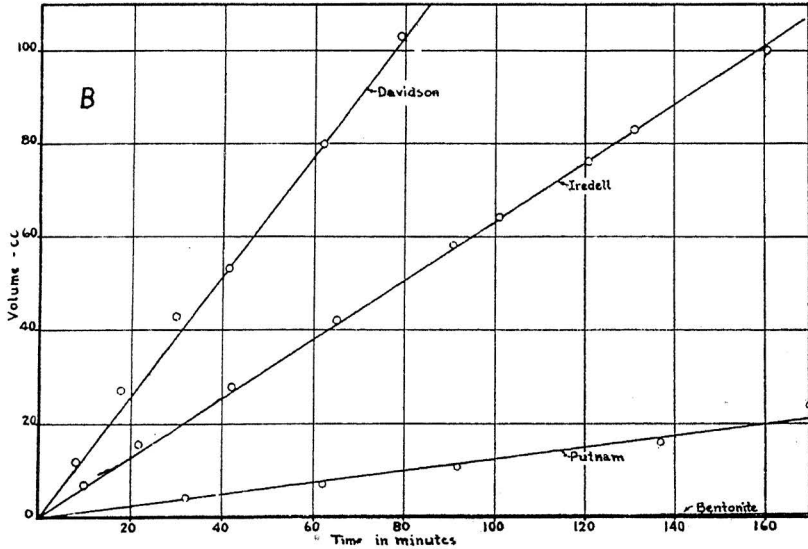
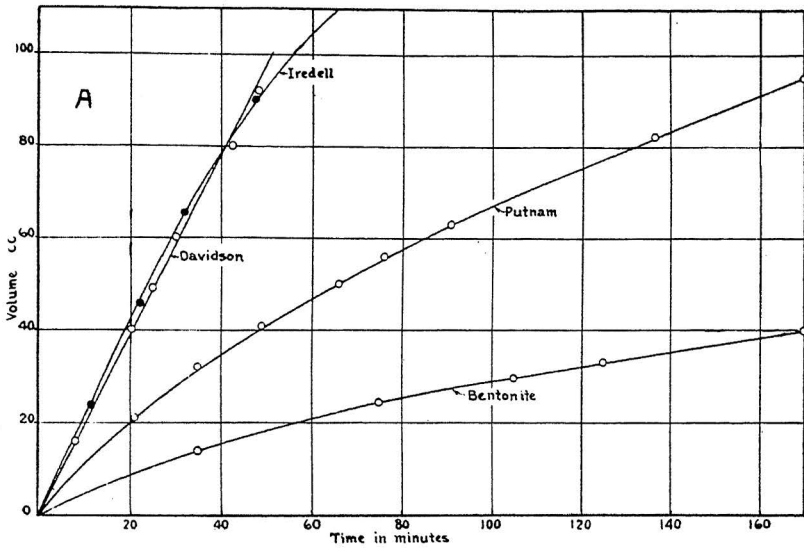


Fig. 9.—The permeability of different clays saturated with Ca-ions. (A) ultrafiltration of sol, (B) permeability of clay membrane.

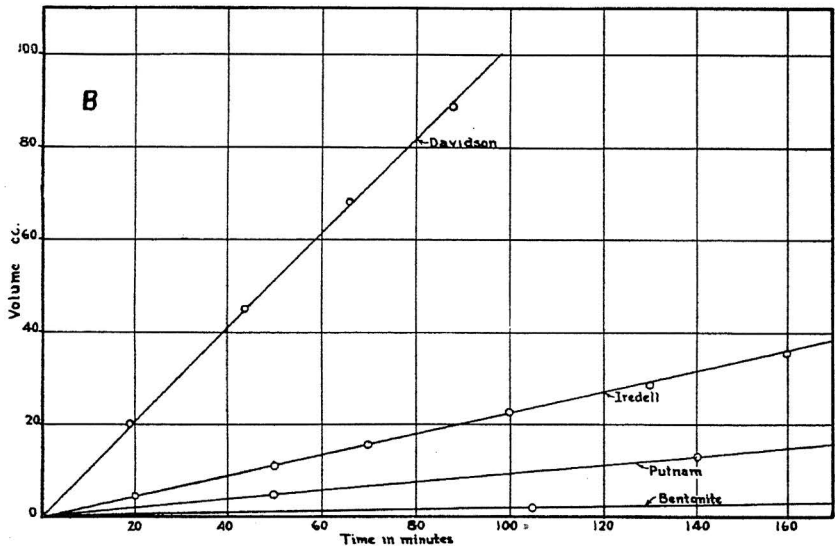
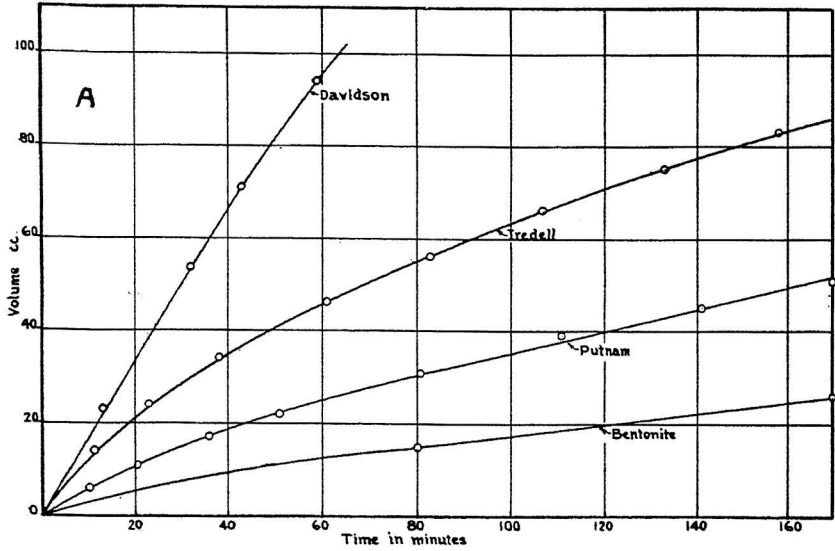


Fig. 10.—The permeability of different clays saturated with Na-ions. (A) ultrafiltration of sol, (B) permeability of clay membrane.

This order is just the reverse of the swelling and of the $\text{SiO}_2\text{-R}_2\text{O}_3$ ratios of these colloids. These data point out quite distinctly that the nature of the inner layer has a marked effect upon the hydration of the colloidal particle. The effect of the exchangeable cations appears to be only of secondary importance.

The non-hydrated granular conditions of the Na-Davidson clay membrane is shown in Figure 11, as compared with the Na-saturated Putnam and Iredell membranes. This photograph was taken soon after removing the membranes from the ultrafilter. It is further evidence of the capacity of the Davidson colloid to resist hydration and to form stable granules in the soil. Studies on the colloidal material extracted from the non-erosive Davidson and the erosive Iredell soils have shown that the hydration and state of dispersion of the colloids are important factors in determining the differences in the erosiveness of the two soils. The resistance of the Davidson clay to erosion can be attributed to the non-hydrated and granular condition of the colloidal fraction. Consequently, rapid percolation of water through the profile is permitted which in turn reduces the amount of surface run-off and erosion. The relative rate of percolation through the Davidson and Iredell clay membranes is conclusive evidence of the greater permeability of the Davidson soil. The dense, impermeable membranes formed by the Iredell colloid indicate quite clearly that the permeability of the B horizon of this soil is due to the hydrated and deflocculated state of the colloids present. The investigation of the stability, hydration and permeability of the colloids extracted from these soils has made possible a clearer concept of the effect of the physical properties of soils on erosion.

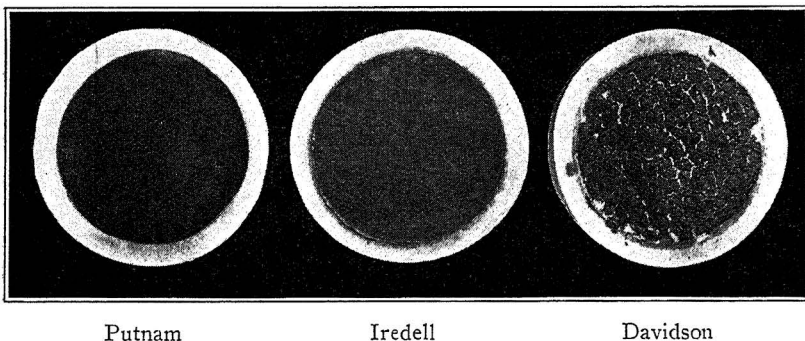


Fig. 11.—Photographs of different Na-clay membranes.

DISCUSSION

Field observations show that soils exhibit vastly different erosive properties even under the same meteorological and environmental conditions such as rainfall, vegetation and topography; the Iredell and Davidson soils of North Carolina are characteristic examples of such differences; the Iredell is a very erosive and the Davidson a non-erosive soil. The B- horizon of the Davidson as it occurs in the field always appears friable, granular, porous and permeable to water, whereas the B-layer of the Iredell is distinctly plastic, sticky and impervious. Inasmuch as the clay content of the Davidson is higher than the Iredell, the differences in erosiveness can not be explained on the basis of texture; the mechanical composition of the two soils should favor higher erosion in the Davidson.

Laboratory studies of the physical properties of these soils conclusively show that one of the principal differences between the Davidson and Iredell from the standpoint of soil erosion is the extent to which the soils are aggregated. The Davidson contains a large amount of aggregates practically all of which are larger than .25 mm. in diameter; the Iredell contains a somewhat lower quantity most of which are relatively small compared with those of the Davidson. Microscopic examinations have shown that the Davidson aggregates are composed of clusters of smaller granules and therefore are permeable, whereas the Iredell aggregates are dense and impermeable. The greater permeability of the Davidson profile is the direct result of the higher content of large aggregates, which are more friable, porous and stable than those of the Iredell. The greater stability of the Davidson aggregates is indicated by the extent to which the soil is naturally granulated and by the resistance of the soil to dispersion as evidenced by a lower dispersion ratio.

Studies with the extracted colloids show that the Davidson is flocculated when saturated with H-, Ba-, Ca-, Li-, Na- or K- ions; additions of NaOH up to 150 per cent of the saturation capacity failed to produce dispersion. On the other hand, only the H-, Ba-, and Ca- Iredell colloids were flocculated. The state of flocculation of the colloids is evidence of the high state of aggregation of the natural Davidson soil.

Measurements of the migration velocities showed that no correlation existed between the $\text{SiO}_2\text{-R}_2\text{O}_3$ ratio and the zeta-potential of the particles as calculated by the Helmholtz-Perrin equa-

tion. Possible explanations of these unexpected data are found, first, in the greater dissociation of the Davidson and Iredell clay acids as indicated by the symmetry values and potentiometric measurements of the "active" H- ions, and secondly, in the fact that the Davidson and Iredell colloids are less hydrated than the Putnam. The degree of hydration would be expected to influence the value of the electrokinetic potential since the Helmholtz-Perrin equation does not take into account differences in hydration. Particles with no water shell would be expected to move faster per "active point" on the complex than those with a thick water shell.

The data on swelling show that the Davidson colloids do not swell irrespective of the nature of the exchangeable cations present; the Putnam and Iredell colloids, however, become hydrated. The Putnam colloid is more highly hydrated than the Iredell. The data on the intensity of the charge and hydration of the various clay sols suggest very strongly the conclusion that hydration is a more important factor than charge in determining the stability of clay suspensions.

Viscosity determinations and consequent calculations of the amount of water associated with a gram of colloid show that the Iredell is less hydrated than the Putnam. A large quantity of water was associated with the various Davidson systems but it is reasonable to assume that most of this water is occluded between the flocculated particles since the Davidson does not swell and all of its suspensions rapidly flocculate. Determinations of the swelling of the different clay colloids showed that the Li- and Na- saturated clays exhibited the highest volume change. The K- saturated clays showed the least amount of swelling. The increase in volume of the Ca-, Ba-, and H- clays was in the order named. Swelling of the latter systems is inversely related to their viscosities and to their relative flocculation times. Apparently, differences in the extent of aggregation of the H-, Ba-, and Ca clays has considerable influence on the degree of swelling.

Permeability measurements show the Li-, Na-, and K- clays less permeable than those saturated with H-, Ba-, and Ca- ions. The lower permeabilities of the former are undoubtedly due to their higher state of dispersion and hydration (except the K- clays). The H- saturated clays are more permeable than the Ba- and Ca- systems which, according to swelling and viscosity measurements, is apparently due to a lower hydration of the H- saturated colloids. Since the H- clay membranes were more permeable in all

cases than the Ca- membranes, and, since the Ca- clays swelled as much or more than the H- clays, it is questionable whether additions of lime to acid soils result in a direct beneficial effect upon the structural condition of the soil. The beneficial effects are probably indirect because of an increase in the organic matter content of the soil, which results in the formation of more stable granules. Displacement of K- and Na- ions from some soils by calcium should result in a more highly flocculated and permeable system. The data on hydration and flocculation show that replacing Na-ions from a colloid by Ca-ions should result in a dehydration, and flocculation of the system; replacement of the K- ions should cause a greater flocculation but very little change in hydration.

Results obtained from the aforementioned studies of the colloidal fraction extracted from the Davidson and Iredell colloids conclusively show that the Davidson colloid is highly flocculated, non-hydrated and extremely permeable to water. The Iredell colloid, on the other hand, is only flocculated when saturated with H-, Ba-, and Ca- ions, is rather appreciably hydrated and is impermeable to water. These data explain the differences in the state of aggregation and the permeability of the two soils. The erosiveness of the Iredell, therefore, must be due chiefly to the ease with which it is dispersed and its impermeability to water. The non-erosive nature of the Davidson is due to the high state of flocculation of its finer mechanical separates into large, porous and stable granules which resist dispersion and permit a rapid percolation of water through the soil profile. The physical properties of soils which affect permeability and the ease of dispersion of the soil particles are paramount factors determining the erosiveness of various soils.

SUMMARY

1. The non-erosive nature of the Davidson clay is largely due to the high degree of aggregation of the B- horizon into large, porous and stable granules.

2. The erosive character of the Iredell is due to its ease of dispersion and the dense, impervious nature of the B- horizon. A small content of aggregates $> .25$ mm. in diameter is present in this layer.

3. All Davidson colloids were flocculated irrespective of the nature of the exchangeable cation on the complex. Only the H-, Ba- and Ca- saturated Iredell colloids were flocculated.

4. The electrokinetic potentials of the colloids from the Iredell, Davidson and Putnam soils were of practically the same value, indicating that hydration rather than charge was the main contributing factor to stability in those systems.

5. The order of swelling for different colloids was bentonite > Putnam > Iredell > Davidson. This is the reverse order of the SiO_2 - R_2O_3 ratios.

6. Exchangeable Ca-, Ba-, and H- cations decreased the swelling of the colloids in the order named. The K-, Na- and Li- cations showed no definite order of effect on the swelling of colloids.

7. The order of permeability of the different clay membranes was H > Ba > Ca > K > Na > Li.

8. The permeability of the H-, Ba- and Ca- membranes was in the same order as the relative viscosity of these systems. The order of permeability was the reverse of that for swelling.

9. The high state of aggregation of the Davidson soil is due to the flocculated and non-hydrated condition of the colloidal fraction.

10. Hydration of the Iredell colloid is responsible for the low state of aggregation and impermeability of the natural soil.

11. Physical properties of soils affecting permeability and the ease of dispersion are paramount factors influencing the erosiveness of various soils.

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