DIELECTRIC DISPERSION METHOD FOR NON-DESTRUCTIVE QUANTIFICATION OF SOIL COMPOSITION

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ABSTRACT

The Maxwell-Wagner-Fricke theory for electrical conduction through heterogeneous media has been extended to develop a theoretical relationship for fluid-saturated porous media, accounting for particle orientation, porosity and the dielectric constants of the pore fluid, solid phase and mixture. This theory is further extended to account for the dielectric dispersion behavior of clay-water-electrolyte systems by considering the surface conductance of particles. The variation of the dielectric constant, ε, and the conductivity, σ, as a function of frequency is the electrical dispersion behavior. The magnitude of the variation of ε over the radio frequency range (1-100 MHz) is referred to as $\Delta\varepsilon_0$. Dielectric dispersion behavior is shown to be controlled by the mineral-solution interface characteristics as influenced by mineralogy and the pore fluid composition. $\Delta\varepsilon_0$ is shown to be related to swell potential and compression index. The electrical dispersion method presented, provides a rapid non-destructive method of quantifying the composition of soils.

KEYWORDS

Dielectric constant, dielectric dispersion, mineral composition, mineral solution interface characteristics, swell potential, compression index, surface conductance, soil classification

INTRODUCTION

In geotechnical engineering, soil classification is used for soil identification, estimation of properties, and translation of experience. By 1908, Atterberg had developed a classification of size fractions based on decimal multiples of 2μ and 6μ, with "clay fraction" defined as the percentage by weight of particles smaller than 2μ. However he realized that particle size alone provided an insufficient basis for the classification of cohesive soils, and decided that a measure of their "plasticity" in terms of liquid and plasticity limits provided to a large extent the additional criteria required. With some modification in nomenclature, this system has been used in geotechnical engineering for the

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past 80 years.

Difficulties with the existing systems of soil classification, as well as the considerable variability between these systems, led to a critical review by Casagrande (1948) and the proposal of the Unified Soil Classification system, adopted in 1952 by the U.S. Corps of Engineers and Bureau of Reclamation, and subsequently by many other organizations (USBR 1963). The Unified Soil Classification System, like all other procedures based on grain size and the properties of remolded materials, can not always reflect the compositional characteristics of a soil, nor can it account for soil structure, state, ambient conditions, or other factors which determine the specific values of the various soil properties. Additionally, the commonly used boundary between particle sizes of non-clay minerals and clay minerals of 2 μm may lead to incorrect classifications because, in natural soils, there are ordinarily particles of clay minerals that are larger than 2 μm in size as well as non-clay particles smaller than this size. Therefore, an arbitrary division based on grain size serves only as a rough measure in determining soil composition.

In a series of five papers, Lambe and Martin (1953-1957) reported compositional data for a large number of soils, then discussed the relationship between composition and engineering properties. These studies showed that the plasticity characteristics of soils containing mixtures of clay minerals such as found in natural soils may be less than would be predicted using knowledge of the properties of the pure clay minerals present and the percentage of each of the clay minerals in the soil as shown by the data in Figure 1 (Lambe & Martin, 1957). In addition, the percentage of clay mineral by weight, as determined by x-ray diffraction or differential thermal analysis, can be more or less than the percentage clay size (2 μm), as shown by the data in Figure 2, (Lambe and Martin, 1957, and Basu and Arulanandan, 1973). Aggregation, cementation and inter stratification of clay minerals are usually cited as the reasons for clay mineral contents in excess of the clay size content.

Atterberg Limits are often unable to properly estimate soil properties. For two soil samples with the same gradation, both containing a small percentage of fines (e.g. 5%), it

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would not be possible to determine Atterberg Limits for either sample. However, if the fines in one sample contained clay minerals, and the fines in the other contained non-clay minerals such as silica flour, the mechanical properties of the two soils would be different. Lambe and Martin found that Atterberg Limits could not always accurately evaluate compressibility. For soils indigenous to St. Louis and elsewhere they found that "very large changes in volume can occur with changes in load or changes in moisture conditions and that these changes are partially reversible. The Atterberg Limits of the St. Louis soil (Sample 19, [L.L.] = 58, [P.L.] = 22) do not suggest such unusual compressibility characteristics" (Lambe and Martin, 1957).

The parameters used to identify swell potential, activity and percent finer than 2µm, are not adequate for accurate evaluation of percent swell for natural soils. Seed, et al., (1962) correlated activity to percent finer than 2 µm to evaluate the swell potential of soils. Their evaluations were reasonably accurate for the artificial soils they tested. A study of both natural and artificial soils made by Basu and Arulanandan (1973) evaluated the ability of the activity index and the percent clay size to evaluatre swell potential. The composition, percent clay size, activity and percent swell for each sample is tabulated in Table 1. The percent swell for the samples was then plotted on the chart developed by Seed, et al., (1962), Figure 3. It can be seen that the percentage total swell predicted for samples 1PB, 3SC and Marysville Red (MR) is significantly lower than the measured value. Thus, while the predicted value is reasonable for artificial soils, the predictions are inconsistent for natural soils. The soil samples from Table 1 are also plotted in Skempton's (1953) activity chart, Figure 4. The three solid lines represent the three clay minerals montmorillonite, illite and kaolinite, from highest to lowest activity. These lines were developed by mixing quartz sand with varying percentages of the three clay minerals. Comparison of the samples to the solid lines shows that activity does not consistently follow mineralogy. For example, Soil 15 with 15% montmorillonite, is expected to plot near the solid line representing montmorillonite with an activity of 7.2. Soil 15 actually plots below the line

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representing kaolinite (Activity = 0.38), suggesting an entirely different mineralogy. Note that samples 1PB, 3SC, 11, 9, 3 and 8 show similar deviations. Thus activity is not an accurate predictor of soil mineralogy.

In the construction of a section of road in Marysville, California, the subgrade consisted of a silty soil with the following properties: L.L. = 46, P.I. = 11, percent less than 2μ = 20, percent less than 1μ = 14, percent sand size = 10, percent silt size = 70, and percent clay size = 20 (Arulanandan, 1974). Three different methods were used to evaluate the swell potential: Skempton's Activity (1953), Holtz and Gibbs percent of free swell (1956), and a chart developed by Seed et al. (1962). All of these methods predicted that the soil would have a low swell potential. However, during construction, the desired density could not be obtained regardless of the compactive effort due to the expansive nature of the soil. In order to determine the actual swell potential of the soil, free swell tests were conducted in the laboratory. Numerous samples of the soil were compacted then subjected to swell at various water contents and dry densities, in half inch deep by four inch diameter Teflon lined rings with porous plugs. The free swell of the samples varied up to 30% within the small range of water contents. This suggested that the soil should instead be considered to have a high swell potential (Arulanandan, 1974).

The previous examples have made several points about some of the existing methods used in conventional geotechnical analysis: (1) gradation analysis alone is insufficient to characterize soils unless the mineralogy of the fines is taken into consideration, (2) Atterberg Limits alone are insufficient for evaluating the volume change characteristics of some natural soils (Lambe and Martin, 1957), and (3) the percent less than 2μ and the activity criteria are also insufficient for evaluating swell potential of some natural soils.

Mineralogy is the primary factor controlling the sizes, shapes, and surface characteristics of the particles in a soil. The type of minerals and the interactions between the solid and fluid phases determine the swelling, compression, strength, and permeability

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of soils. Thus mineralogy and pore fluid composition are fundamental factors controlling geotechnical properties.

Arulanandan et. al, 1973, proposed a non-destructive method to characterize the type and amount of clay minerals and the mineral-solution-interface characteristics (electrical double layer) using electrical methods. The objective of this paper is to demonstrate the significance of the magnitude of dielectric dispersion in the radio frequency range for the non-destructive characterization of soil composition.

RADIO FREQUENCY ELECTRICAL DISPERSION AND SOIL COMPOSITION

The application of an electrical impulse in the form of a voltage to any geological material in the radiofrequency range produces a response which can be measured in terms of the dielectric constant ε and conductivity σ. In non-clay minerals, the ε and σ response is independent of frequency as demonstrated in Figure 5, whereas in hydrated soils with clay minerals, the ε and σ vary as a function of frequency as also shown in Figure 5. Figure 6 shows the variation of the dielectric and conductivity behavior (electrical dispersions) of saturated Marysville soil which contains clay minerals. The variation of ε or σ with frequency is called the electrical dispersion. The magnitude of dielectric dispersion, $\Delta\varepsilon_0$, which is denoted in Figure 5 and also shown for the Marysville soil in Figure 6, is defined as the change in ε measured over the radio-frequency range. The magnitude of dielectric dispersion is determined by using the high and low frequencies at which the dielectric constants in the dispersion curve level off. The magnitude of dielectric dispersion of saturated soils has been shown to be a function of the mineralogy and the mineral-solution interface characteristics (due to double layer polarization). Therefore, the magnitude of dielectric dispersion can be used to quantify the composition of soils. Arulanandan et. al (1973) have proposed the dielectric dispersion method to characterize the type and amount of clay minerals in soils.

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It can be seen from the dielectric dispersion behavior of most natural soils as shown in Figure 7, (Sargunam, 1973), that if the dielectric measurements are made at 2 MHz and 50 MHz, the magnitude of dielectric dispersion, $\Delta\varepsilon_0$ can be obtained. For soils which have dielectric dispersions which extend beyond the range of 2 and 50 MHz, such as the Marysville soil, $\Delta\varepsilon_0$ may be extrapolated from the experimental data using a Cole-Cole plot (Arulanandan, 1966) as was done in Figure 6 for the Marysville soil. A theoretical basis for this new index, $\Delta\varepsilon_0$, and its use to characterize the mineralogy and the mineral-solution interface characteristics as influenced by the pore fluid composition and particle shape is presented herein. The applicability of the dielectric dispersion behavior of saturated soils for the evaluation of swell potential, and compression index is also examined.

ELECTRICAL CONDUCTION THROUGH HETEROGENEOUS MEDIA

Arulanandan, et al., (1985) extended Maxwell (1881) and Fricke's (1924, 1953) analysis of the electrical properties of suspended ellipsoidal particles in an alternating current field to provide a theoretical basis to describe the electrical dispersion behavior of transversely isotropic soils, and then applied the theory to the evaluation of porosity (Arulanandan, et al., 1985; Arulanandan, 1991). The theory developed to describe the electrical dispersion behavior is summarized below.

Maxwell (1881) derived the following expression for the conductivity of a heterogeneous media consisting of spherical particles in a dilute electrolyte suspension, assuming that the electric field of one particle does not influence the electric field of another:

$$\frac{k}{k_1} = \frac{2k_1 + k_2 - 2(1-n)(k_1 - k_2)}{2k_1 + k_2 + (1-n)(k_1 - k_2)}$$

where k is the conductivity of the medium, k_1 is the conductivity of the solution, k_2 is the conductivity of the particle, and n is the porosity. Wagner (1914) extended Maxwell's equation to consider the relaxation mechanisms of the dielectric constant and conductivity that occurs with heterogeneous media under alternating current which is generally referred

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to as the Maxwell-Wagner relaxation mechanism. In clay soils, charges accumulate at the interface (mineral solution interface, the electrical double layer) between the clay particle and the surrounding solution. Since this build-up of charges takes time, as the frequency is increased there will be less time for the charges to accumulate at the interface, which in turn decreases the system's ability to store electrical potential energy, and thus decreases the dielectric constant. When the frequency reaches a certain value, there will not be enough time for charges to accumulate at the interface, and at this point, the dielectric constant becomes independent of frequency. The value of the dielectric constant at this leveling-off frequency is defined as ε_∞.

Fricke (1953) considered the particles to be ellipsoid, with all particles oriented in one direction only in dilute suspensions. Results of the theoretical dielectric dispersions considered in this paper show that Fricke's theory is valid for higher volume concentrations. Dafalias and Arulanandan (1979, 1983) considered ellipsoid particles of any axial-ratio oriented in multiple directions based on a probability distribution of orientations.

Consider the soil particles to be spheroids with semi-axes a, b and c, as in Figure 8, where b equals c and b and c are not equal to "a" in general. When $b/a \geq 1$, the particles are modeled as oblate spheroids and when $b/a < 1$, the particles are modeled as prolate spheroids. The following analysis holds for both oblate and prolate spheroids.

Representing the multiple orientation of particles by appropriate probability density functions, as proposed by Dafalias and Arulanandan (1979, 1983), expressions were obtained for the vertical (F_v) and horizontal (F_h) formation factors as a function of porosity, n, as follows (Dafalias and Arulanandan, 1979, 1983)

$$F_v = \frac{k_1 - k_2}{k_v - k_2} = 1 + \frac{1-n}{n} f_v \tag{1}$$

$$F_h = \frac{k_1 - k_2}{k_h - k_2} = 1 + \frac{1-n}{n} f_h \tag{2}$$

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where

$$f_v = \frac{P_\theta}{1 + \frac{k_2 - k_1}{k_1} A_a} + \frac{1 - P_\theta}{1 + \frac{k_2 - k_1}{k_1} A_b} \tag{3}$$

and

$$f_h = \frac{1}{2}\left[\frac{1 - P_\theta}{1 + \frac{k_2 - k_1}{k_1} A_a} + \frac{1 + P_\theta}{1 + \frac{k_2 - k_1}{k_1} A_b}\right] \tag{4}$$

In the above expressions k_1, k_2 are the complex electrical conductivities of the solution and the particle, respectively. k_v and k_h are the complex electrical conductivities of the composite medium in the vertical direction and the horizontal direction, respectively. Complex conductivity k is defined as

$$k = \sigma + j\omega\varepsilon \tag{5}$$

where σ is the conductivity, ε is the apparent dielectric constant, ω is the angular frequency and $j = \sqrt{-1}$. ω is related to the frequency, f, by the expression $\omega = 2\pi f$. The shape indices, A_b and A_a for a spheroid are defined as (Fricke, 1953):

$$A_b = \frac{1}{2\left(1 - R^2\right)}\left[\frac{R^2}{2\sqrt{1 - R^2}}\ln\left(\frac{1 - \sqrt{1 - R^2}}{1 + \sqrt{1 - R^2}}\right) + 1\right], \quad \text{for } 0 \le R \le 1 \tag{6a}$$

and

$$A_a = \frac{1}{2\sqrt{R^2 - 1}}\left[\frac{R^2}{\sqrt{R^2 - 1}}\tan^{-1}\sqrt{R^2 - 1} - 1\right], \quad \text{for } R > 1 \tag{6b}$$

and $A_a = 1 - 2A_b$, where axial ratio, $R = b/a$ (Fricke, 1953). Some values of shape indices for special cases are given: for spherical particles $R = 1$ and $A_b = 1/3$, for laminae shaped particles $R \to \infty$ and $A_b \to 0$, and for long infinite cylinders $R = 0$ and $A_b = 1/2$. The orientation factor, P_θ, in eqs. (3) and (4) represents the multiple orientations of particles and is defined as

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$$P_\theta = \int_0^{\pi/2} p(\theta)\cos^2\theta\, d\theta \tag{7}$$

where the probability density function, $p(\theta)$, characterizes the distribution of the orientation of semi axis "a" with respect to the vertical direction for $0 \leq \theta \leq \pi/2$ such that (Dafalias and Arulanandan, 1979)

$$\int_0^{\pi/2} p(\theta)\, d\theta = 1 \tag{8}$$

An explicit expression for k_v can be obtained by rearranging eq. (1) as

$$k_v = k_1 + (k_2 - k_1)\,\frac{(1-n)f_v}{n+(1-n)f_v} \tag{9a}$$

A similar expression for k_h can also be obtained by rearranging eq. (2) as

$$k_h = k_1 + (k_2 - k_1)\,\frac{(1-n)f_h}{n+(1-n)f_h} \tag{9b}$$

Equations (9a and 9b) give expressions for complex electrical conductivities in the vertical and horizontal directions k_v and k_h, respectively, in terms of the complex conductivities of the solution, k_1, the particles, k_2, porosity, n, the particle orientation, P_θ, and the axial ratio, R. When each of the complex conductivities in eqns. (9a and 9b) are replaced by the conductivities and dielectric constants as given in eq. (5) the following equations are obtained:

$$\sigma_V + j\omega\varepsilon_V = \sigma_1 + j\omega\varepsilon_1 + \frac{[\sigma_2 - \sigma_1 + j\omega(\varepsilon_2 - \varepsilon_1)](1-n)f_V}{n + (1-n)f_V} \tag{10a}$$

$$\sigma_h + j\omega\varepsilon_h = \sigma_1 + j\omega\varepsilon_1 + \frac{[\sigma_2 - \sigma_1 + j\omega(\varepsilon_2 - \varepsilon_1)](1-n)f_h}{n + (1-n)f_h} \tag{10b}$$

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where:

$$f_v = \frac{P_\theta}{1+\dfrac{\sigma_2-\sigma_1+j\omega(\varepsilon_2-\varepsilon_1)}{\sigma_1+j\omega\varepsilon_1}A_a} + \frac{1+P_\theta}{1+\dfrac{\sigma_2-\sigma_1+j\omega(\varepsilon_2-\varepsilon_1)}{\sigma_1+j\omega\varepsilon_1}A_b} \qquad (11a)$$

$$f_h = \frac{1}{2}\left\{\frac{1-P_\theta}{1+\dfrac{\sigma_2-\sigma_1+j\omega(\varepsilon_2-\varepsilon_1)}{\sigma_1+j\omega\varepsilon_1}A_a} + \frac{1+P_\theta}{1+\dfrac{\sigma_2-\sigma_1+j\omega(\varepsilon_2-\varepsilon_1)}{\sigma_1+j\omega\varepsilon_1}A_b}\right\} \qquad (11b)$$

σ_v and σ_h are the conductivities of the system in the vertical and horizontal directions respectively, σ_1 and σ_2 are the conductivities of the solution and the particles respectively, and ε_1 and ε_2 are the dielectric constants of the solution and particles respectively.

Thus the real part of eqns. (10a and 10b) will give expressions for the variation of vertical conductivity and horizontal conductivities, σ_v and σ_h , as a function of ω, σ_1, ε_1, σ_2, ε_2, P_θ, R, and porosity n. The imaginary part of eqns. (10a and 10b) will give expressions for the variation of the vertical and horizontal dielectric constants of the system, ε_v and ε_h , as a function of ω, σ_1, ε_1, σ_2, ε_2, P_θ, R, and n. The dielectric constant, ε_1, is approximately equal to 80. The measured value of the dielectric constant of dry particles, ε_2, is about 4.5 (Arulanandan, 1973).

CONDUCTIVITY OF THE SOIL PARTICLE σ_2

The electrical conductance of a soil particle σ_2 depends on the mineralogy and the adsorbed pore fluid. The electrical conductivity of the particle may be defined as the sum of two conductivities. The first is due to the conductivity of the charged carriers of the particle (conductivity of the spheroid) and the second conductivity is due to the existence of the double layer in the interface between the clay particle and the surrounding pore fluid (surface conductance). For dry particles the double layer is not present and the conductivity of the particle is due only to the charges of the particle. For saturated systems,

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