

Induced polarization of soils and unconsolidated sediments

J. Richard,¹ A. Ghorbani,¹ A. Revil,² P.S. Pauw,³ G-J. Nijsten,³ J. R. Delsman,³ R. Harting,⁴ J. L. Gunnink,⁴ J. Monney² and P. Vaudelet¹

¹*Naga Geophysics, F-73000 Chambéry, France*

²*Université Savoie Mont-Blanc, CNRS, UMR CNRS 5204, EDYTEM, F-73370 Le Bourget du Lac, France. E-mail: andre.revil@univ-smb.fr*

³*Deltares, Daltonlaan 600, 3584 BK Utrecht, P.O. Box 85467, 3508 AL Utrecht, the Netherlands*

⁴*TNO—Geological Survey of the Netherlands, Princetonlaan 6, PO Box 80015, NL-3508 TA Utrecht, the Netherlands*

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SUMMARY

Induced polarization can be used to image the water content and Cation Exchange Capacity (CEC) of soils with applications in agriculture and soil sciences. This can be done thanks to a robust petrophysical model that can be used to interpret field data. Here we use two petrophysical models based on the classical Dynamic Stern Layer (DSL) polarization model. The first was developed using a volume averaging approach. A second DSL model was recently developed based on fractal theory. Furthermore, we include the presence of semiconductors (pyrite or magnetite) in these two models. In order to test their predictions, the complex conductivity spectra of 400 samples of shallow sediments and soils were measured in the frequency range 1 Hz–45 kHz at 25°C and at three different NaCl solutions. The two models predict that the surface conductivity and the normalized chargeability are proportional to each other and depend on the CEC and water content. The experimental data set favours the modified DSL model based on the fractal approach. The effects of the CEC and volume fraction of semiconductors are investigated since framboidal pyrite is present in the investigated coastal soils and sediments from the Netherlands. Experiments show that the amount of semiconductors can lead to an apparent cementation exponent below unity in agreement with the petrophysical models. This observation is further and qualitatively confirmed by additional measurements obtained by mixing a soil with different amounts of magnetite. The presence of the organic matter (OM) changes the CEC of the material acting therefore on both the surface conductivity and normalized chargeability. Our analysis shows that pure OM has a CEC of 160 ± 40 meq per 100 g. An application of the volume-averaged-based DSL model to a field study in Guadeloupe is performed to showcase how we can map the water content and CEC of soils at a field scale using time-domain induced polarization measurements. The novelty of the analysis lies into the relationship between the normalized chargeability measured with the time-domain induced polarization method (injecting the current for a given period) and the normalized chargeability obtained in the frequency-domain and equal to the difference between the instantaneous conductivity and the DC conductivity. The application of the petrophysical models to field data implies that the amplification factor between the two normalized chargeability can be calibrated using experimental data or determined through numerical modelling by integrating frequency-domain spectra into a time-domain response accounting for the duration of the input current.

Key words: Electrical properties; Permeability and porosity; Induced polarization; Hydrogeophysics.

1 INTRODUCTION

Soils and unconsolidated sediments are a fundamental component of the critical zone of the Earth, which is sustaining all

terrestrial life (A.-L. FAO 2023). Their proper management is therefore a key component to the long-term development of our societies on this planet (J.L. Havlin 2020). Intrusive and non-intrusive methods can be used to image and monitor their

properties. Therefore they should not be overlooked (A. Romero-Ruiz et al. 2019).

Geophysical methods are widely used in mapping properties of the subsurface of the Earth. Electrical resistivity tomography (ERT) is a broadly used, non-intrusive technique used to image the electrical conductivity of the subsurface of the Earth at various scales. However, quantitative ERT is often challenging without additional measurements as it depends on two contributions. The first contribution is associated with the pore water conductivity (bulk conductivity) and second contribution (surface conductivity) to the electrical double layer coating clay minerals (M.H. Waxman & L.J.M. Smits 1968; J.D. Rhoades et al. 1976, 1989; F.D. Börner et al. 1996; C.A. Dias 2000; G. Okay et al. 2013). A geophysical method known as induced polarization (e.g. G. Olhoeft 1985; J. Schön 1996; T. Dahlin 2014) can be used to separate these two contributions (see A. Weller et al. 2013; A. Revil 2013). Therefore, induced polarization can be used to monitor the evolution of key textural parameters of unconsolidated sediments as well as environmental changes such as temperature or salinity fluctuations (C.A. Dias 2000; G. Okay et al. 2014; S. Johansson et al. 2016; A. Revil et al. 2017; S. Hupfer et al. 2017) and the Cation Exchange Capacity (CEC) of clays and organic matter (N. Schwartz & A. Furman 2015). This is especially true in the Netherlands where extensive land use and use of the subsurface and groundwater, combined with changing climate, land subsidence and societal needs necessitate a detailed understanding of the architecture and properties of the subsurface as well as the groundwater salinity in the interest of groundwater management and water availability (J.R. Delsman et al. 2018, 2023). Regarding the effect of the Organic Matter (OM), two papers have recently underlined the role of the OM and its aromatic content on the quadrature conductivity spectra (F. Delarue et al. 2025; T. Bore et al. 2026). That being said, their analysis were not yet rooted in firm mechanistic models and are dataset-specific.

A. Revil et al. (2017) developed the first induced polarization model for unconsolidated sediments and soils based on the Dynamic Stern Layer (DSL) concept (O. Stern 1924; H.J. Vinegar & M.H. Waxman 1984). They used a volume-averaging approach to solve the problem. A. Revil et al. (2017) used a data set of 71 samples collected from the Province of Zeeland in the Netherlands to test their model (see also A. Revil et al. 2021, for a field application to a vineyard in France). The results obtained by A. Revil et al. (2017) were later used by J.R. Delsman et al. (2018) in interpreting airborne frequency-domain electromagnetic (FEM) data to infer the 3-D groundwater salinity distribution in Zeeland (in the Netherlands).

Since the publication of these studies, several papers have appeared presenting additional data sets of the complex conductivity of unconsolidated sediments, rocks and soils (see recently L. Piolat et al. 2025; C. Moser et al. 2025) which have confirmed the results of A. Revil et al. (2017). The real part of the complex conductivity corresponds to conduction, that is, electromigration of the electrical charge carriers, through the porous material while the imaginary part (quadrature component) characterizes low-frequency polarization processes associated with the reversible charge accumulation/depletion in the porous material in response to the applied electrical field (G. Olhoeft 1985). The polarization of the porous material should be understood at the grain and pore scales and is connected to the properties of the electrical double layer. C. Strobel et al. (2023) emphasized that organic matter (OM) characterized by a high CEC

(see also A. Revil et al. 2013, 2017) can dominate the imaginary conductivity response of soils and unconsolidated clay-rich sediments. They recommended adding Total Organic Content (TOC) measurements to field data but their study was limited to several tens of core samples and no model was provided. K. Tsukanov & N. Schwartz (2020) emphasized the role of roots in the complex conductivity spectra. S. Saneiyan et al. (2023) pointed out the role of expansion/shrinkage in the determination of the electrical conductivity of expansive soils with smectite (see recently J. El Alam et al. 2024, 2025). S. Saneiyan et al. (2019, 2021), Y. Song et al. (2022) and C. Strobel et al. (2023) emphasized the role of bacterial communities in soils and shallow unconsolidated sediments in their spectral induced polarization characteristics.

In this paper, we use an unprecedented data base of 400 samples of unconsolidated sediments from the southern, western and northern Netherlands. Together, they cover most widely distributed geological units present in the upper 50 m below surface of present day to Oligocene age, although some samples were collected at greater depth. They encompass a broad range of sedimentary environments, that is, aeolian, fluvial, marine, glacial and local deposits. Due to the large area where samples were taken and the large depth range of up to 236 m below the ground surface, they also contain a wide range of salinities present in the Dutch subsurface. Of the 400 samples, 86 were taken as undisturbed *in situ* samples preserving their composition and sedimentary structure with respect to their original texture in the field (undisturbed sampling). The other 314 samples were disturbed samples, as was the case of the study of A. Revil et al. (2017). Three additional samples from Guadeloupe were taken as disturbed samples from a clay-rich nitisil. The total data set includes clay-rich and silty sediments, sandy clays, fine and coarse sands, as well as samples with a high percentage of organic matter including peat. Some were taken from anthropogenically disturbed sediment and soils, mainly from agricultural fields.

The complex conductivity measurements are carried out at three salinities. Porosity and CEC are also independently measured. The CEC is a classical property of soils in agriculture (D.S. Ross 1995; C.E. Johnson 2002). This data set adds to previous studies focusing on different geological settings, for example on low-porosity shales and high porosity bentonites (A. Revil et al. 2023a, b; J. El Alam et al. 2024a, b) including the effect of pyrite (A. Revil et al. 2024). Since the investigated unconsolidated sediments from the Netherlands are demonstrated to contain framboidal pyrite and pyritized plants (E.C.L. Marnette et al. 1993; J. Griffioen et al. 2016; J.E. Vermaat et al. 2016), the effect of pyrite will be investigated into later in this study, which was not taken into account in the earlier study on unconsolidated sediment from the Netherlands by A. Revil et al. (2017). That said, Y. Kai et al. (2025) have recently reported the effect of pyrite on the induced polarization properties of soils and sediments. In our case, the presence of pyrite is confirmed through SEM (Scanning Electron Microscope) pictures from both the literature as well as using a national Dutch reference data set with geochemical data for the same geological units sampled in this study.

The paper is divided into four sections. The first section reviews the classical DSL model based on volume-averaging theory but unable to account for the fractal-characteristics of soils and sediments. An updated DSL model for the complex conductivity of soils and sediments was recently developed by S. Qiang

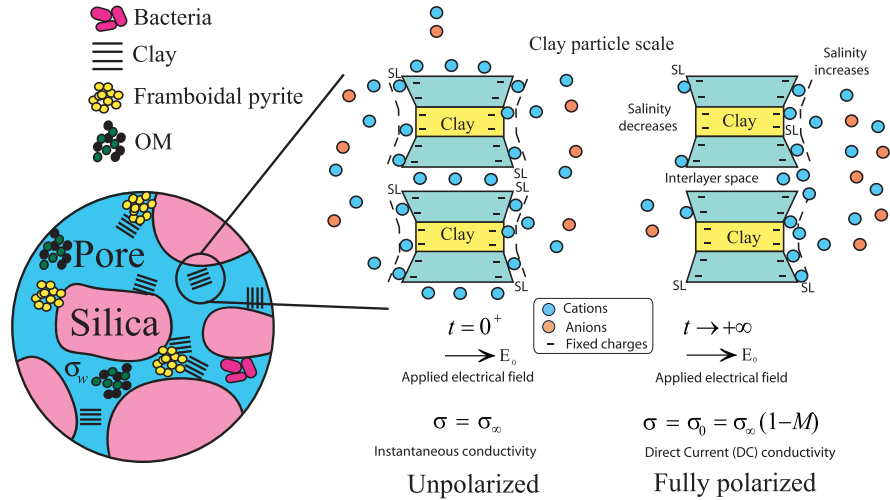


Figure 1. Soil and shallow sediment electrical properties. Such materials are a composite of different elements including coarse grains (fine and coarse silica sands in our case), clay minerals, bacteria and Organic Matter (OM). Each element is characterized by its own electrical properties. The polarization of a clay particle involves electromigration and diffusion of the charge carriers (cations and anions) in the local electrical field resulting from the macroscopic applied (external) electrical field. The conductivity just after the application of this field is called the instantaneous conductivity σ_∞ while the conductivity obtained in the steady-state condition (particle fully polarized) is called the DC (Direct Current) electrical conductivity σ_0 .

et al. (2025). Furthermore, we also discuss the effect of the presence of semiconductors like pyrite (or magnetite) in these two DSL models. In the second section, we discuss the laboratory experiments and the complex conductivity spectra of the samples investigated in this study. We then follow with a comparison between the model and the experimental data to check the efficiency of the updated DSL model to capture the experimental trends. Finally, a field application is presented to showcase how a CEC map can be produced in field conditions using time-domain induced polarization measurements. In doing so, we will emphasize the role of an amplification factor a in transferring the field-acquired normalized chargeability obtained with time-domain induced polarization (current injected during a period T) and the normalized chargeabilities obtained in the laboratory using frequency-domain induced polarization spectra and defined as the difference between the instantaneous conductivity and the Direct Current (DC) conductivity.

2 MODEL

2.1 Soils and unconsolidated sediments without metallic particles

Unconsolidated sediments are complex porous media with a variety of minerals (silica sands and carbonates, clay minerals), with organic matter, microcolonies of bacteria and roots (Fig. 1). The main elements of polarization in sediments remain however the clay minerals and the organic matter because of their large CECs (see Fig. 1 for the polarization of a clay particle). The equivalent circuit of sediment without metallic particles such as magnetite or pyrite (Fig. 2a) is associated with the existence of the electrical double layer coating the surface of the grain (Fig. 2b). As discussed later in this paper, framboidal pyrite (with submicron-sized crystals) is ubiquitous in organic matter-rich sediments and soils in coastal areas (D.T. Rickard 1970; R.T. Wilkin *et al.* 1996). J.R. Vallentyne (1963) reported a density of 10^5 pyrite framboids per gram of dry sediment in the

organic matter of a lake in North America. The formation of these framboids can take just few days in the right redox conditions (D. Rickard 1970).

The expression ‘Dynamic Stern Layer (DSL) model’ implies that the counterions in the Stern layer (in the inner part of the electrical double layer) are mobile under a gradient in the electrochemical potentials of the ionic species in the pore water. The DSL model also implies that the equivalent circuit of a sediment can be represented by a resistor in parallel with another resistor. These two resistors represent the bulk conduction (electromigration of the charge carriers in the external electrical field) through the pore space and the effect of the conduction in the electrical diffuse layer (the external part of the double layer) coating the grains, especially coating clay particles. These conduction paths are in parallel with the frequency-dependent conductivity of the Stern layer represented by a resistance in series with a Warburg (leaking) capacitance (E. Warburg 1899, see Figs 1 and 2).

In the approach developed in this paper, the overlapping of the electrical diffuse layers impedes, to some extent, their polarization. This is in contrast with colloidal particles in the so-called dilute limit for which the solid particles coated with their electrical double layer are immersed in an electrolyte (see J. Lyklema *et al.* 1983; V.N. Shilov *et al.* 2001; P. Leroy *et al.* 2017; M. Bückner *et al.* 2019; I. Markova *et al.* 2024). Then the poly-dispersivity of the grains and pores provides a distribution of relaxation times that is broader than the one expected in the simple Warburg model shown in Fig. 2. Eventually, the distribution of the polarization length scales can obey a power-law distribution characteristic of fractal systems and a specific DSL fractal model was recently developed by S. Qiang *et al.* (2025).

We consider below complex conductivity spectra above 1 Hz corresponding mostly to the polarization of the clay minerals (the sand grains, $>100 \mu\text{m}$, polarize at lower frequencies, $<1 \text{ Hz}$, because of their large sizes with respect to the clay particles, $<1 \mu\text{m}$, and the silt at intermediate frequencies, above 100 Hz) with a component associated with the organic

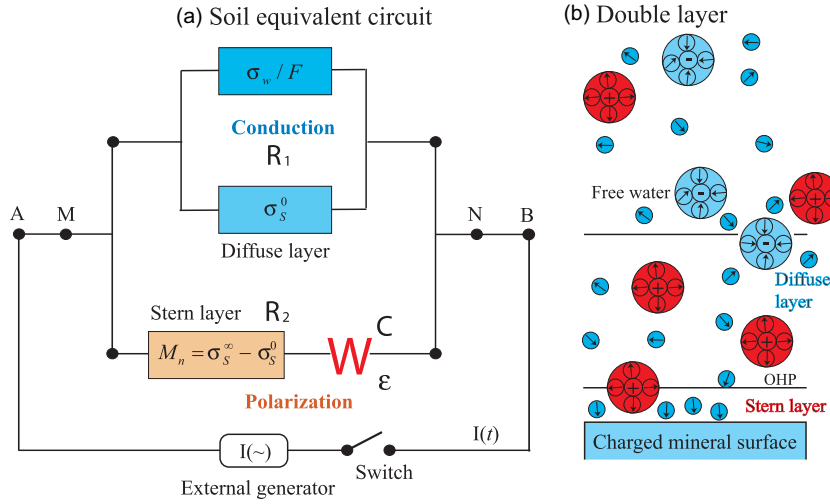


Figure 2. Role of the electrical double layer in the equivalent circuit of a soil or a shallow unconsolidated sediment. (a) Equivalent circuit with AB the current electrodes and MN the voltage electrodes. The normalized chargeability M_n (in $S\ m^{-1}$) is defined as the difference between the instantaneous conductivity and the DC conductivity. It is also equal to the difference between the instantaneous surface conductivity σ_s^∞ and the surface DC conductivity σ_s^0 . This difference corresponds to the Stern layer contribution while the diffuse layer contribution to the surface conductivity (σ_s^0) acts in parallel with the conduction in the bulk pore space σ_w/F . (b) The electrical double layer comprises the Stern and diffuse layers. OHP denotes the Outer Helmholtz Plane at the interface between the two layers.

matter. Furthermore in clay-rich media, because of the high conductivity of the materials, the phase angle is very small below 1 Hz and difficult to get with large measurement errors. The effect of organic matter will be discussed extensively below. The sediment samples are considered to be isotropic and time-invariant, with a connected pore space. Their complex conductivity is written as,

$$\frac{1}{\rho^*} = \sigma^* = |\sigma| e^{i\varphi} = \sigma'(\omega) - i\sigma''(\omega), \quad (1)$$

where ρ^* denotes the complex resistivity of the material. In eq. (1), $|\sigma|$ and φ denote the amplitude (in $S\ m^{-1}$) and phase shift (in rad) of the complex conductivity σ^* , respectively. They can be related to the amplitude and phase of the impedance Z^* (ohm) measured by the impedance meter. The quantities σ' and σ'' (both in $S\ m^{-1}$) denote the in-phase (real) and quadrature (negative, imaginary) conductivity respectively, $\omega = 2\pi f$ is the angular frequency (rad s^{-1}), f denotes the frequency (in Hz) and i denotes the pure imaginary number with $i^2 = -1$. With the sign convention in eq. (1), the quadrature conductivity σ'' is definite positive.

If we consider the circuit shown in Fig. 2(a) (C.A. Dias 1968, 1972; W.H. Pelton 1977), the complex conductivity is given by (C.A. Dias 1968, 1972; W.H. Pelton 1977),

$$\sigma^*(\omega) = \sigma_\infty \left(1 - \frac{M}{1 + (i\omega\tau)^{1/2}} \right), \quad (2)$$

where $\rho_0 = 1/\sigma_0$ corresponds to the DC (Direct Current) resistivity (σ_0 denotes the DC conductivity), the chargeability M (dimensionless) and the relaxation time τ (in s) are given by

$$M \equiv \frac{\rho_0 - \rho_\infty}{\rho_0} = \frac{\sigma_\infty - \sigma_0}{\sigma_\infty}, \quad (3)$$

$$\tau = \frac{r^2}{2D}, \quad (4)$$

and $\rho_\infty = 1/\sigma_\infty$ (in ohm m) corresponds to the instantaneous resistivity (σ_∞ denotes the instantaneous conductivity in $S\ m^{-1}$), r (in m) denotes a characteristic polarization length scale and D ($m^2\ s^{-1}$) is the diffusion coefficient of the counterions when the primary current is shut down and the charges carriers come back to their equilibrium state. In the following, we will use the normalized chargeability M_n (in $S\ m^{-1}$), defined as $M_n \equiv \sigma_\infty M = \sigma_\infty - \sigma_0$, as characterizing the ability of a material to reversibly store electrical charges.

As briefly mentioned above, the Warburg model cannot represent the equivalent circuit of a porous sediment due to its heterogeneous composition. Indeed, a porous sediment is characterized not just by a single polarization length scale, but by a broad distribution of polarization length scales as discussed further below. The previous model needs therefore to be generalized to (A. Revil *et al.* 2017, 2021)

$$\sigma^*(\omega) = \sigma_\infty - M_n \int_0^\infty \frac{h(\tau)}{1 + (i\omega\tau)^{1/2}} d\tau, \quad (5)$$

where $h(\tau)$ denotes a (normalized) probability density for the relaxation times τ (in s) of the material and is called the kernel of the relaxation time distribution (see for instance A. Tarasov & K. Titov 2007). A special case of kernel is the Cole-Cole model.

For saturated sediments, the classical DSL model based on a volume-averaging approach yields the following expressions (A. Revil 2012):

$$\sigma_\infty = \frac{1}{F} \sigma_w + \frac{1}{F\phi} \rho_g B\ CEC, \quad (6)$$

$$\sigma_0 = \frac{1}{F} \sigma_w + \frac{1}{F\phi} \rho_g (B - \lambda)\ CEC, \quad (7)$$

$$M_n = \frac{1}{F\phi} \rho_g \lambda\ CEC, \quad (8)$$

where σ_w (S m⁻¹) denotes the pore water solution conductivity, ϕ the (connected) porosity, F is the formation factor (dimensionless) and ρ_g denotes the grain density (2770±30 kg m⁻³ for clay minerals, 2650±50 kg m⁻³ for silica sand, 2710 kg m⁻³ for calcite). Note that glauconite, a mineral broadly present in several geological units of Pliocene and Miocene age in the Netherlands, has a grain density between 2500 to 2950 kg m⁻³ (depending on the amount of iron) and can therefore substantially increase the grain mass density of the sand. The mass density of organic matter and humus can be in the range 500–600 kg m⁻³ for humus to 1000–1500 kg m⁻³ for organic matter (e.g. M.A. Fosberg 1977). In the following, we will use an average value of 800±300 kg m⁻³ for the organic matter (see A. Revil *et al.* 2017).

The CEC denotes the cation exchange capacity of the porous material (in C kg⁻¹ with 1 meq per 100 g = 963.20 C kg⁻¹ in SI units, Y. Bruneel *et al.* 2021). B (in m²s⁻¹V⁻¹) denotes the apparent mobility of the counterions for surface conduction and λ (in m²s⁻¹V⁻¹) denotes the apparent mobility of the counterions for the polarization associated with the quadrature conductivity (see H.J. Vinegar & M.H. Waxman 1984). We have B (Na⁺, 25°C) = $3.1 \pm 0.3 \times 10^{-9}$ m²s⁻¹V⁻¹ and λ (Na⁺, 25°C) = $3.0 \pm 0.7 \times 10^{-10}$ m²s⁻¹V⁻¹ (A. Revil *et al.* 2023a, b; 2024; A. Ghorbani *et al.* 2023; and references therein). In contrast to the empirical model of H.J. Vinegar & M.H. Waxman (1984), the model of A. Revil *et al.* (2017) connects both B (in m²s⁻¹V⁻¹) and λ to the true mobilities of the counterions in the Stern and diffuse layer and to the fraction of counterions in the Stern layer. Furthermore, while the model of H.J. Vinegar & M.H. Waxman (1984) is empirical and introduces, in an *ad hoc* way, a porosity term in the quadrature conductivity, the model of Revil and co-workers (A. Revil 2012, 2013; A. Revil *et al.* 2017) is based on volume-averaging approach and account for the frequency dependence of the spectra.

A dimensionless number R was introduced by A. Revil *et al.* (2017) as $R = \lambda/B \approx 0.10 \pm 0.02$ (see updates in A. Revil *et al.* 2023a, b, 2024; A. Ghorbani *et al.* 2023) independent of temperature and saturation. This ratio is also equal to the ratio between the instantaneous surface conductivity (the last term of eq. 3, i.e. $\sigma_s^\infty = \phi^{m-1} \rho_g B \text{CEC}$) and the normalized chargeability $M_n = \phi^{m-1} \rho_g \lambda \text{CEC}$ (A. Revil 2012). This explains the proportionality between the normalized chargeability and the surface conductivity as observed, for instance, by F. Börner (1992), A. Revil (2013), A. Weller *et al.* (2013) and D. Emmanuel *et al.* (2024).

A different DSL model corresponds to an extension of the fractal-based model developed by H. Wang & A. Revil (2020) for the electrical conductivity. It was recently developed by S. Qiang *et al.* (2025). Unconsolidated sediments exhibit complex microstructures with length scales varying over several orders of magnitude, from nanometres to millimetres. Polarization length scales (grains or pores) follow, at the first order, power laws (A. Ougier-Simonin *et al.* 2016; P. Cosenza *et al.* 2021). This hierarchical structure may often be characterized by the fractal dimension D_f (C.E. Krohn 1988; H. Van Damme 1995; A.-L. Fauchille 2015). The S. Qiang *et al.* (2025) DSL model yields the following expression for the instantaneous conductivity, the DC conductivity and the normalized chargeability,

$$\sigma_\infty = \phi^m \sigma_w + B[\rho_g(1 - \phi) \text{CEC}]^{m_s}, \quad (9)$$

$$\sigma_0 = \phi^m \sigma_w + (B - \lambda)[\rho_g(1 - \phi) \text{CEC}]^{m_s}, \quad (10)$$

$$M_n = \lambda[\rho_g(1 - \phi) \text{CEC}]^{m_s}, \quad (11)$$

where m and m_s denote the cementation (porosity) exponent and a new exponent defining an interfacial Archie's law [m and m_s are related to the fractal dimension D_f of the pore space of the porous material by $m_s = m + (m - 1)/(3 - D_f - 1)$; H. Wang & A. Revil 2020]. So computing from the data set the two exponents m and m_s allows to determine the fractal dimension of the pore space. The quantity $\rho_g(1 - \phi) \text{CEC}$ represents the excess charge per unit total volume (and not per pore volume as in the models of M.H. Waxman & L.J.M. Smits 1968, and H.J. Vinegar & M.H. Waxman 1984).

Because sediments exhibit complex microstructural structures with length scales varying over several orders of magnitude, from nanometres to millimetres, the distribution of the relaxation times in the kernel $h(\tau)$ (see eq. 2) of the complex conductivity can be very broad resulting in a constant phase model. Assuming such behaviour, and considering the quadrature conductivity at the geometric mean frequency of two frequencies f_1 and f_2 and the normalized chargeability defined as the difference between the in-phase conductivity at the frequency $f_2(> f_1)$ and the in-phase conductivity at the lower frequency f_1 , the quadrature conductivity and the normalized chargeability can be related to each other by (A. Revil *et al.* 2017, their Appendix A),

$$\sigma''(\sqrt{f_1 f_2}) \approx -\frac{M_n(f_1, f_2)}{\alpha}, \quad (12)$$

where $\alpha \approx (2/\pi) \ln A$ with A represents the number of decades between the two frequencies f_1 and f_2 (the ratio between the two frequencies). This relationship works for intermediate frequencies since the polarization length scales cannot be smaller than a certain size (typically 0.1 μm) or larger than the biggest pore or grain size (typically 100 μm) (H.J. Vinegar & M.H. Waxman 1984). Therefore, three orders of magnitude in grain/pore sizes would translate into 6 orders of magnitude in relaxation times. In other words, the constant phase model may be valid over six decades of frequencies, possibly 0.1 Hz to 100 kHz. If a sediment would have a more limited variation in grain sizes, for example, in very homogenous sand or a clay with an exceptionally large clay fraction, the quadrature conductivity spectra would exhibit well-defined polarization peaks. In this paper, eqs (9) to (12) provided an alternative DSL model called the DSL-fractal model.

2.2 The effects of the organic matter

The two previous petrophysical models (DSL volume-averaging and DSL-fractal models) require the CEC as an input parameter. In sediments and soils, the CEC is a composite of the CEC associated with the clay minerals and the CEC associated with the organic matter (e.g. N. Schwartz & A. Furman 2015). Considering the excess of charge per unit volume of material, we have $\rho_g(1 - \phi) \text{CEC} = \varphi_C \rho_g^C \text{CEC}_C + \rho_g^{OM} \varphi_{OM} \text{CEC}_{OM}$ where C refers to the clay minerals, OM refers to the Organic Matter, ρ_g^C and ρ_g^{OM} refer to the mass density of the clay minerals and organic matter (without its microporosity), respectively, and φ_C and φ_{OM} refer to the volume fraction of clay and organic matter, respectively. The TOC can be measured by different methods such as Loss On Ignition (LOI), the Automated Dry Combustion (ADC) standard method, and Humic Matter (HM) colorimetry (W.R.

Roper *et al.* 2019). The CEC of the organic matter is typically in the range 150–400 meq per 100 g, which is larger than the CEC of smectite, the clay mineral characterized by the highest CEC (typically 100 ± 20 meq per 100 g, e.g. C. Pan & Y.-H. Shen 2003).

2.3 The effects of metallic particles in soils and shallow sediments

The last step of our model is to include the effect of the volume-content of metallic particles to the previous petrophysical models. Indeed, sediment from the Netherlands may contain some semiconductors like pyrite or magnetite, as well as an occasionally high amount of glauconite (up to more than 50 per cent in weight fraction in some sands from the Netherlands). Therefore, we pursue, like in A. Revil *et al.* (2024), our goal to see the effect of such semiconductors on the effective properties of such mixtures. We know that the presence of semiconductors may have a strong influence on the induced polarization properties of rocks and sediments. As shown by A. Revil *et al.* (2015a, b) the presence of pyrite in proportions with less than 20 per cent in volume content modifies the conductivity and chargeability (dimensionless but often expressed in mV V⁻¹) as,

$$\sigma_{\infty} = \left(\frac{1}{F} \sigma_w + \sigma_s^{\infty} \right) (1 + 3\varphi_m), \quad (13)$$

$$\sigma_0 = \left(\frac{1}{F} \sigma_w + \sigma_s^0 \right) \left(1 - \frac{3}{2} \varphi_m \right), \quad (14)$$

$$M \equiv \frac{\sigma_{\infty} - \sigma_0}{\sigma_{\infty}} \approx \frac{R}{1 + \frac{1}{F} \frac{\sigma_w}{\sigma_s^{\infty}}} + \frac{9}{2} \varphi_m, \quad (15)$$

where φ_m denotes the volume fraction of metals, semimetals and semiconductors in the porous material. Eqs (13) to (15) are very versatile since the instantaneous and DC surface conductivity models are given by, respectively, the last terms of eqs (6) and (7) for the DSL volume averaging-based model and by the last terms of eqs (9) and (10) for the DSL fractal-based model. We can get a new expression of the normalized chargeability as

$$M_n \equiv \sigma_{\infty} M, \quad (16)$$

$$M_n = R \sigma_s^{\infty} (1 + 3\varphi_m) + \frac{9}{2} \varphi_m \sigma_{\infty}, \quad (17)$$

which update the previous two petrophysical models by explicitly adding the effect of the presence of the metallic particles.

In order to better interpret laboratory measurements, we introduce an apparent formation factor (F_a), an apparent surface conductivity (σ_s^a) and an apparent normalized chargeability (M_n^a) to obtain an expression similar in form to the expressions without pyrite (or any metallic particles) (see the so-called Archie's law, G.E. Archie, 1942):

$$\frac{1}{F_a} \equiv \phi^{m_a} = \phi^m (1 + 3\varphi_m), \quad (18)$$

$$\sigma_s^a = \sigma_s^{\infty} (1 + 3\varphi_m), \quad (19)$$

$$M_n^a = R \sigma_s^{\infty} (1 + 3\varphi_m), \quad (20)$$

and therefore,

$$\sigma_{\infty} = \frac{1}{F_a} \sigma_w + \sigma_s^a, \quad (21)$$

$$M_n = M_n^a + \frac{9}{2} \varphi_m \sigma_{\infty}, \quad (22)$$

where m_a (dimensionless) denotes an apparent cementation (porosity) exponent in Archie's law (G.E. Archie, 1942). The presence of metals, semimetals and semiconductors therefore results in (1) a decrease of the apparent formation factor and apparent cementation exponent with respect to the background value (i.e. the sediment without the presence of a semiconductor like pyrite), (2) an increase of the apparent surface conductivity and (3) a strong increase of the normalized chargeability. Coming back to the cementation exponent m , we can provide a relationship describing the decrease of the apparent cementation exponent with the increase of the metal content φ_m . Starting with eq. (18), we obtain after few algebraic manipulations the following relationship,

$$m_a = m + \gamma, \quad (23)$$

where the correction factor γ is given by,

$$\gamma \equiv \frac{\ln(1 + 3\varphi_m)}{\ln \phi} \leq 1. \quad (24)$$

Taking $\varphi_m = 0.15$ (15 per cent volume content of pyrite, for instance) and a porosity $\phi = 0.60$ yields $\gamma = -0.73$. Taking $m = 1.4$ yields $m_a \approx 0.7 < 1$. Therefore, the apparent cementation exponent of sediment rich in organic matter and therefore in framboidal pyrite can be smaller than one. We will see that this is indeed the case in our experiments.

3 MATERIALS AND METHODS

3.1 Samples

In this study, we first use 400 samples from shallow unconsolidated sediment and soils from the Netherlands (Fig. 3). Three samples from Guadeloupe Island will be discussed later in the application section and 12 additional experiments were performed with a sandy soil mixed with magnetite. We first focus below on the Dutch samples.

In 2022 and 2023, a drilling and sampling campaign was undertaken to acquire samples from the southern, western and northern parts of the Netherlands (Fig. 3). The locations were chosen to collect a representative set of samples for processing geophysical data from an anticipated Airborne Electromagnetic (AEM) survey in these portions of the Netherlands. The sampling locations were selected to take into account the geological heterogeneity of the subsurface and variations in groundwater salinity up to a depth of ~ 200 m, roughly corresponding to the AEM survey depth. In total, 43 drillings were used, comprised of (1) manual auger drillings up to 8 m below surface (comparable to the drillings used in A. Revil *et al.* 2017); (2) mechanical suction/air lift drillings up to 236 m below surface, partly with undisturbed cores and (3) mechanical (semi-)continuous core drillings up to 60 m below surface. All drillings were macroscopically described in terms of lithology and subsequently lithostratigraphically interpreted according to the Dutch lithostratigraphic Nomenclator (K. TNO-GDN 2024). The samples were taken from geological units dating from recent anthropogenically influenced to late Oligocene sediments; they were collected from marine, fluvial, aeolian, glacial and organogenic sedimentary environments.

The height of the core samples holder is approximately 5 cm, and its diameter is 4.6 cm. From the undisturbed cores, a total of 86 pairs of samples were collected, each pair containing an undisturbed sample and a disturbed, scooped, sample. These

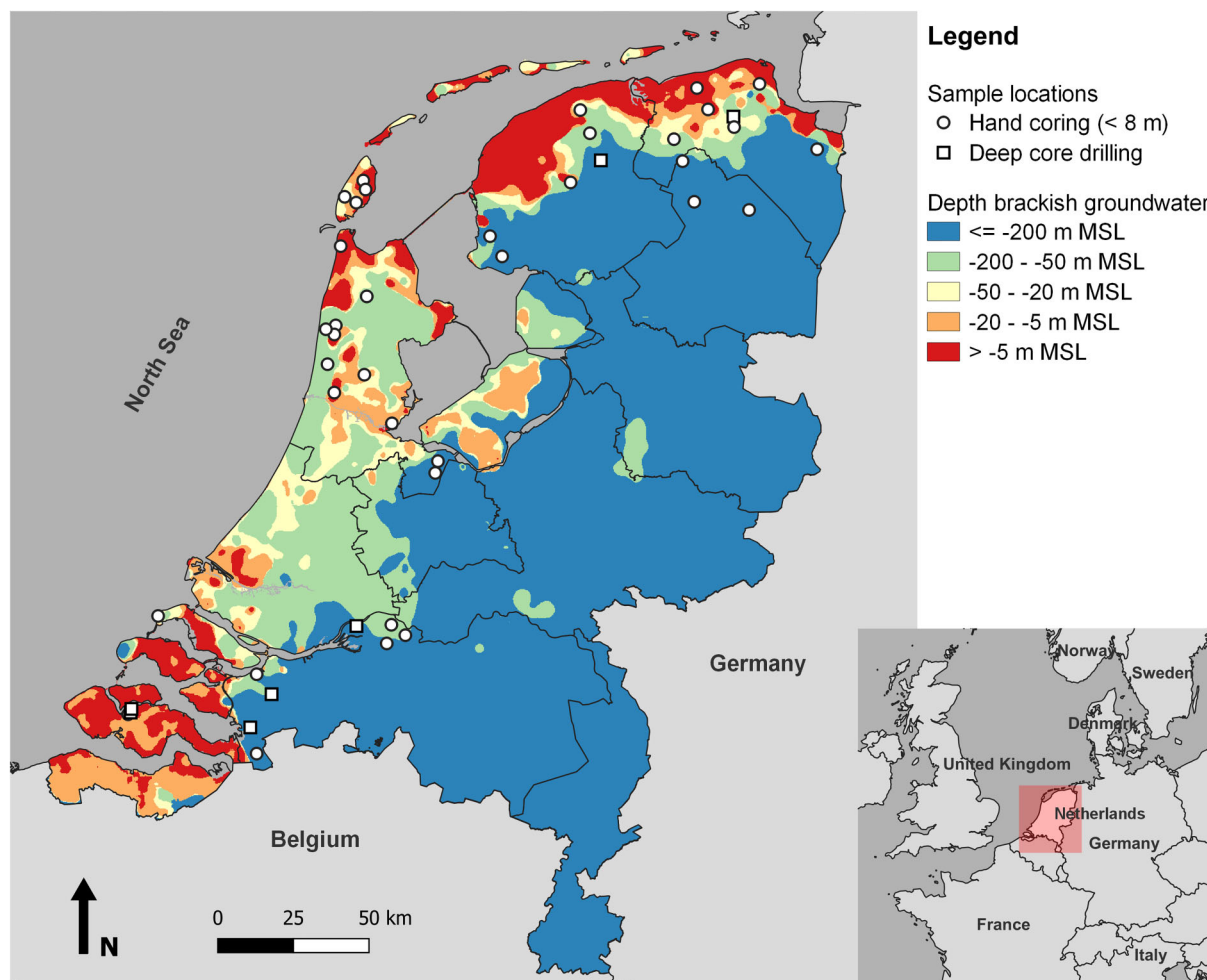


Figure 3. Sample locations and modelled depth of the interface between brackish and saline groundwater below mean sea level (MSL, see J.R. Delsman *et al.* 2023). MSL (Mean Sea Level) represents the average height of the ocean's surface, typically over several years.

pairs were taken from cored intervals with the internal *in situ* structure of the sediment intact. The other 228 samples were taken from disturbed drilling material. Besides the samples used for determining their complex conductivity spectra, pore water was extracted to determine the Electrical Conductivity (EC) value using Rhizon CSS samplers with 0.15 μm pore size (Rhizosphere Research Products, <https://www.rhizosphere.com/>). The pore water was extracted close to the induced polarization sample intervals, usually in the same lithological layer. The pore water electrical conductivity generally varied from fresh water to brackish water (1 S m^{-1} at 25°C), and was used as a reference for the lowest salinity experiments (see below). Regarding the clay mineralogy, both fluvial and marine clays are generally dominated by illite and smectite, whereas kaolinite and chlorite typically only contribute up to 20 per cent of the $<2 \mu\text{m}$ weight fraction. For more information on the mineralogy of the Dutch Cenozoic sediments, the reader is referred to J. Griffioen *et al.* (2016, 2025).

3.2 Pyrite content

Pyrite is known to be present in the Dutch subsurface (e.g. E.C.L. Marnette *et al.* 1993; C. Van Kempen & J. Griffioen 2011;

J. Griffioen *et al.* 2016). Although we did not quantify the volumetric content of pyrite of our samples, we derived a qualitative indicator (classification) based on the existing extensive knowledge derived from the national Dutch reference data set comprising over 5000 geochemical analyses (including total Sulphur, TS and Total Organic Carbon, TOC) with associated lithological and geological information (Vernes *et al.* 2010, and subsequent use of this data set can be found for instance in P.F.M. Van Gaans *et al.* 2011 and J. Griffioen *et al.* 2016).

In general, we can state that the pyrite content of the sediment below a few metres depth and relatively low in organic content (no peat) is generally related to Total Sulfur (TS, see Table 1). Note that such TOC-TS relationships have also been applied and verified in other sediments worldwide (J. Griffioen *et al.* 2016). For organic matter-rich samples, a relatively high pyrite content can be expected, occurring as pyritized plant remains or framboids of 10–50 μm on mostly clay particles but also sand grains (see Fig. 9). Depending on the sedimentary environment, some clays and occasionally organic rich sands may also contain elevated levels of pyrite. In general, sand contains the least pyrite (averages of 0.04 to 0.4 per cent), clays somewhat higher (averages of 0.04–1.26 per cent) and peat the highest (0.45 to 4.95 per cent). In the acknowledgments of this paper, the data used in

Table 1. CEC (cation exchange capacity), TOC (total organic carbon) and pyrite presence for a subset of the cored samples. Class refers to the main lithological class of the samples (V = peat, K = clay, Z = sand). TS denotes the Total Sulfur content. TOC was determined independently by two different labs; EDYTEM (France) and ISOLAB (the Netherlands). They differ in methodology especially in their ignition temperature, but data could be corrected using a linear scaling factor. Uncorrected data are presented here.

ID	Lithology	Class	CEC (meq per 100 g)	TOC(1) (per cent)	TOC(2) (per cent)	TS (2) (per cent)	Presence of pyrite
Fd051	Very clayey peat	K	22.2	6.7	–	–	Likely
Fd052	Very sandy peat	V	36.0	36.7	–	–	Very likely
Fd059	Peat	K	20.3	12.5	8.7	1.70	Very likely
Fd060	Clayey peat	V	186.6	54.3	–	–	Very likely
Fd062	Organic-rich clay	V	6.5	3.2	–	–	Very likely
Fd083	Peat	V	28.8	9.1	–	–	Very likely
Fd086	Peat	V	39.5	70.1	43.8	0.23	Likely
Fd095	Peat	K	22.6	53.5	–	–	Likely
Fd151	Peat	V	169.4	71.6	–	–	Very likely
Fd153	Peat	V	69.3	–	13.4	1.20	Very likely
Fd154	Organic rich clay	V	191.7	–	38.5	3.55	Very likely
Fd178	Organic rich clay	V	94.0	78.2	–	–	Very likely
Fd182	Mod. org. rich clay	V	51.2	–	37.5	1.00	Very likely
Fd186	Peat	K	45.4	16.0	–	–	Very likely
Fd191	Peat	K	5.2	5.1	2.3	0.04	Very likely
Fd195	Slightly clayey peat	K	36.9	–	6.3	0.47	Very likely
Fd205	Peat	V	25.1	–	37.1	4.30	Very likely
Fd206	Peat	V	189.9	66.1	–	–	Very likely
Fd215	Peat	V	21.5	–	31.0	2.10	Very likely
Fd217	Very organic rich	V	25.6	43.0	25.2	4.05	Very likely
Fd223	Moderately fine sand	V	97.8	81.8	46.0	1.05	Very likely
Fd236	Loose grainy peat	V	27.7	–	35.0	4.90	Very likely
Fd237	Mod. silty slightly	Z	11.4	–	0.7	0.15	Likely
Fd258	organic clay	V	88.0	78.7	–	–	Very likely
Fd259	Extr. silty slightly	K	21.4	7.5	–	–	Likely
Fd260	organic clay	K	23.2	7.6	1.1	0.43	Likely
Fd264	Peat	V	10.1	23.3	–	–	Very likely

(1) TOC data by EDYTEM. (2) TOC data by ISOLAB.

this paper can be found, including the value of the pyrite indicator of each sample and a brief motivation for the classes used. We have listed the pyrite content indicator of all samples; this indication is independently assigned based on the rationale above and our knowledge on the lithological, geological and redox conditions of the samples. Table 1 lists a selection of these samples for which the TOC and TS content were also determined.

3.3 Porosity, CEC and TOC measurements

The CEC data were obtained for all 403 collected samples with the cobalt hexamine technique (D. Aran *et al.* 2008). The cobalt is used to perform a titration of the excess charge occurring onto the mineral surface. This titration is performed with a 0.05 N hexamine cobalt (III) chloride solution (40 mL). The colour of cobalt hexamine depends on its concentration so it is easy to determine the amount fixed by the clay minerals and organic content, and therefore the value of the CEC (charge per unit mass of dried material). Sand and silt have a much lower absorption capacity than clays and can be neglected. The organic matter is characterized by very high CEC values. In our case, 5 g of material was used for sand-rich samples and 2 g or less was used for clay- and organic matter-rich samples. After agitation of samples with the cobalt hexamine, the CEC was determined by measuring the difference of absorbance between pure cobalt hexamine chloride solution and solutions mixed and equilibrated with the

powders from the sample. The porosity of the core samples was obtained from the dry weight and saturated weight and the volume of the samples.

The TOC was measured using a first subset 19 samples using the classical Loss On Ignition (LOI) method (e.g. D.W. Nelson & L.E. Sommers 1996) at EDYTEM Laboratory (France). The samples are first dried for 24 hr at 58°C and then incinerated at 550°C for 4 hr. The TOC (fraction or per cent) values are obtained by determining the mass difference as $TOC = 1 - m_1/m_2$ where m_1 denotes the initial dry mass of a sample and m_2 the mass after the LOI process. Additional measurements were performed in the Netherlands (ISOLAB) using a second subset of 14 samples (from the Netherlands) grinding the material to fine powders. For these samples, the drying temperature used in the TOC was higher at 100°C instead of 58°C.

3.4 Complex conductivity measurements

The protocol used for the complex conductivity measurements was similar to that used by A. Revil *et al.* (2017). The spectral induced polarization (SIP) measurements were investigated over the frequency range 1 Hz–45 kHz using the high-precision impedance analyser developed by Egon Zimmermann in Germany (E. Zimmermann *et al.* 2007; 2008) (see Fig. 4). Four equally spaced non-polarizing sintered Ag/AgCl electrodes (4 mm × 1 mm) were placed at the top of the sample holder

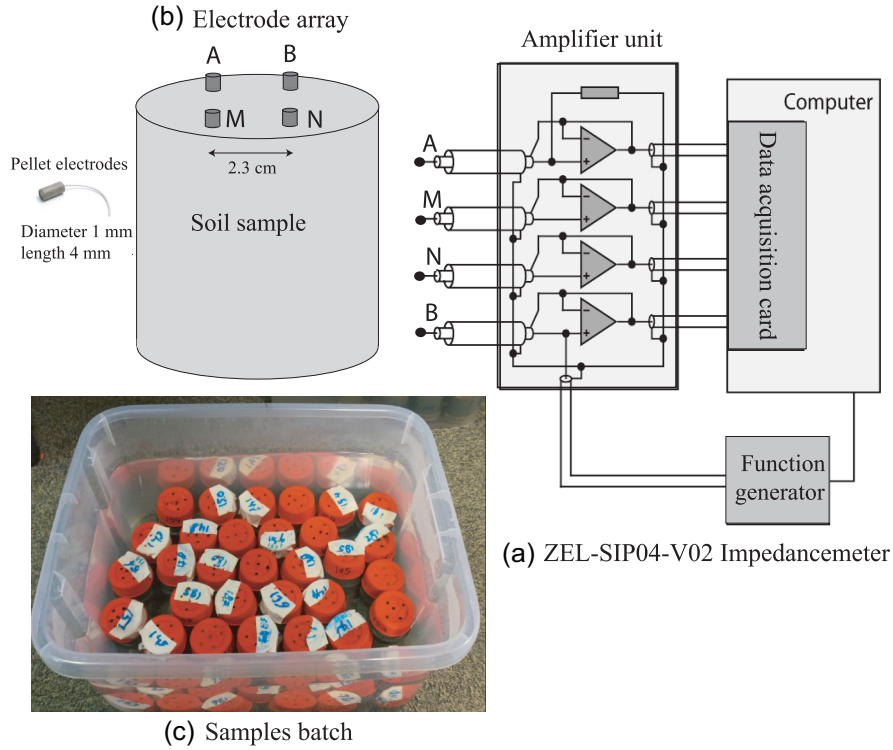


Figure 4. Impedance meter and samples. (a) ZEL-SIP04-V02 impedance meter developed by Egon Zimmermann at Juelich (Germany, see E. Zimmermann *et al.* 2007, 2008). (b) Position of the Ag/AgCl electrodes used for the measurements. The geometrical coefficient used to convert the impedance into complex conductivity is determined by solving the Laplace equation for the potential accounting for the geometry of the sample and the position of the current electrodes A and B and voltage electrodes M and N. (c) Example of a batch of samples. The collection of the core samples was divided into six batches for the measurements with 54 to 86 samples per batch.

(Fig. 4) The interval between two electrodes was 2.3 cm so much larger than the size of the sand grains $< 300 \mu\text{m}$ (see Fig. 4). The set of samples was organized in batches (see Fig. 4) and the measurements were performed over a period of 9 months.

For the first and lowest salinity, the samples were dried in the autoclave at 45°C during 24 hr, then saturated under vacuum with a NaCl solution of a known conductivity. For the coarse sand samples, we used a pore water conductivity solution at 0.03 S m^{-1} (at 25°C). For the other samples, we used a NaCl solution at 0.1 S m^{-1} (at 25°C). The core samples are let to equilibrate at these pore water conductivities for a month. After saturation under vacuum, the samples were submerged in the solution for three weeks to complete the saturation of their pore space. For the second and third water conductivities, the change of brine was done by diffusion by submerging the samples in a large tank (see Fig. 4c) in which the solution is set at the desired salinity. The second and third water conductivities are 0.1 and 1 S m^{-1} (NaCl, 25°C) for the coarse sand samples and 1 and 10 S m^{-1} for the other soil samples. Equilibrium was reached after 4 weeks for both steps. The water conductivity was measured and controlled by a calibrated electrical conductivity meter (wtw 3310, measuring between $0.001 \mu\text{S cm}^{-1}$ to 100 S m^{-1}).

A geometrical factor K is used to convert the impedance to resistivity (or the conductance to conductivity). The sample holder has a height of 5 cm and a diameter of 4.6 cm with an open-top. Given the limited surface area available (only one section),

we use square electrode setup rather than the classical linear array. To determine K , we carried out a test using the sampled holder filled with a solution of known conductivity. The geometrical factor was also computed from numerical modelling using the finite element code COMSOLTM Multiphysics, simulating the current injection/retrieval at electrodes A and B and solving the Laplace equation for the electrical potential with insulating boundary condition at the boundaries of the sample holder except for electrodes A and B (their surface area was accounted for in the computations). Details regarding this procedure can be found in D. Jougnot *et al.* (2010). We found a geometrical factor of $8.6 \pm 0.3 \text{ m}^{-1}$. The error bar given for the geometric factor from the numerical model is due to the uncertainty on the height. Examples of complex conductivity spectra are shown in Fig. 5 for the four groups of lithologies, being fine and coarse sands, clays and organic-rich sediments.

Both the in-phase conductivity and the absolute value of the quadrature conductivity increase with the frequency (Fig. 5). The dispersion (frequency-dependence) of the in-phase conductivity can be used to define the partial normalized chargeability, which is the difference between the in-phase conductivity values at two different frequencies. The electrical conductivity of the material (as well as surface conductivity) is expected to increase monotonically with frequency and reaches two asymptotic limits at low and high frequencies (σ_0 and σ_∞ , respectively). This partial normalized chargeability is necessarily smaller than the normalized chargeability defined as the difference between the

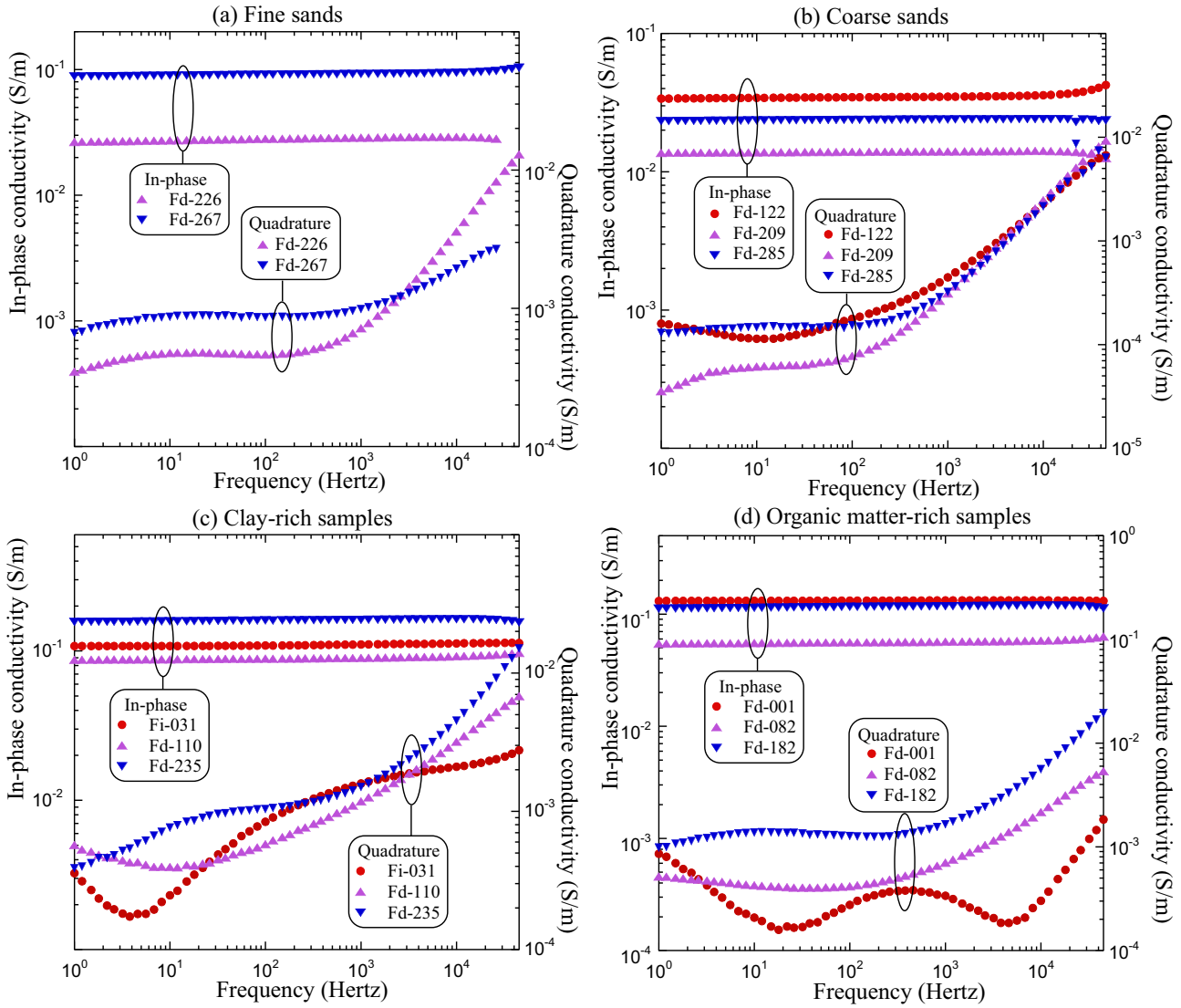


Figure 5. Complex conductivity spectra in the frequency range 1 Hz–45 kHz for the four lithological groups. (a) Fine sands. (b) Coarse sands. (c) Clays. (d) Organic matter rich-sediments and soils (peat). We show the first level of pore water conductivity (0.03 S m^{-1} for coarse sand samples and 0.1 S m^{-1} for the other samples). The strong increase of the quadrature conductivity at high frequencies (higher than 1 or 10 kHz, depending on the samples) is due to the Maxwell-Wagner polarization (generally above 1 to 5 kHz).

instantaneous conductivity (high-frequency) and the DC conductivity (low frequency)

$$M_n(f_1, f_2) \leq M_n(\infty, 0) \equiv \sigma_\infty - \sigma_0, \quad (25)$$

but both can be related to each other (see for instance A. Revil *et al.* 2017). This point is very important to be sure that the normalized chargeability determined between two frequencies or at a given period for the injected current in time-domain induced polarization measurement can be properly related to the total normalized chargeability $M_n(\infty, 0)$ used in the model. The latter [i.e. $M_n(\infty, 0)$] corresponds to the normalized chargeability between the DC condition and the high-frequency (instantaneous) conductivity. This is also the normalized chargeability obtained in time-domain induced polarization if the current is injected for an infinite time $(-\infty, 0)$.

4 RESULTS

4.1 Determination of the formation factor and surface conductivity

Fig. 6 depicts the in-phase conductivity (at 1 Hz) of six samples with different lithologies as a function of the pore water conductivity at the three salinities mentioned in the previous section. The trends are usually analysed with a linear conductivity model $\sigma' = a\sigma_w + b$ where a denotes the inverse of the formation factor and b denotes the surface conductivity σ_s (see eq. 21, a and b are the fitted parameters), not taking the effect of metallic particles like pyrite into account. The fit is performed in a log-log space and the fitting function is therefore,

$$Y = \log(a10^X + b), \quad (26)$$

where $Y = \log \sigma'$ and $X = \log \sigma_w$ and a and b have the same meaning as above and are still the fitted parameters. Once the

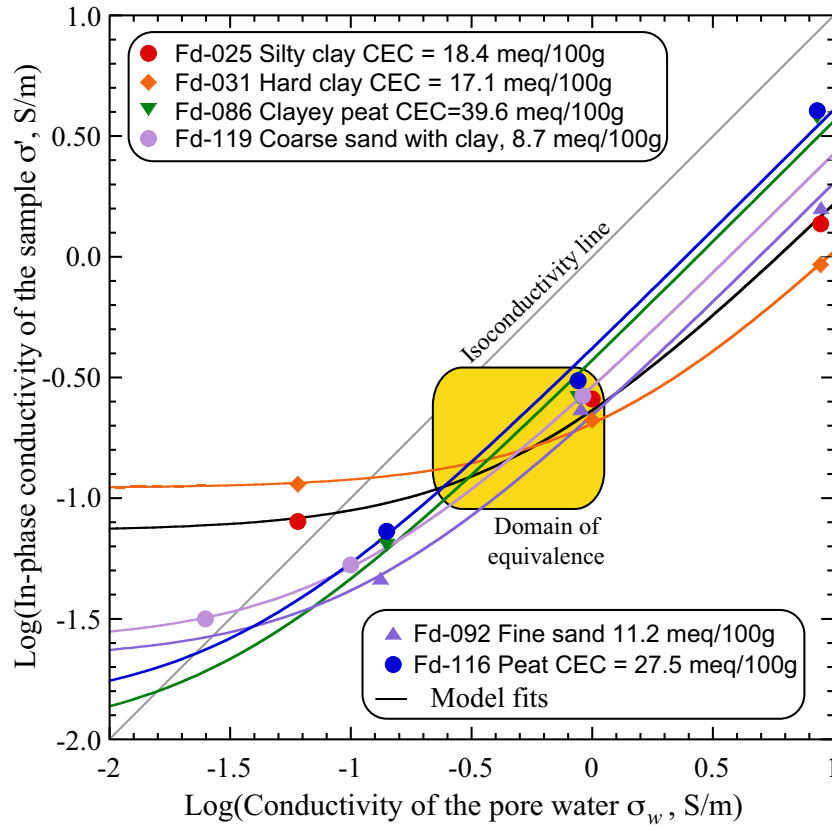


Figure 6. In-phase conductivity of six samples (taken at 1 Hz, 25°C) versus the pore water conductivity. We use the linear conductivity model described in the main text but in a log–log space to determine the intrinsic formation factor F and the surface conductivity σ_s . The isoconductivity line corresponds to the line for which the conductivity of the material is equal to the conductivity of the pore water and equal to the surface conductivity. The contoured area corresponds to a pore water conductivity for which all the materials have roughly the same conductivity (conductivity alone cannot be considered as a lithological indicator). Note that the use of three salinities is the bare minimum to be confident in the determination of both the surface conductivity and formation factor.

formation factor and the surface conductivity are determined for each sample, we can try to correlate the formation factor to the porosity and the surface conductivity to CEC. Coming back to Fig. 6, we observe that at low salinities, the clay-rich and organic matter-rich samples are more conductive than the sandy-soils (see V. Emelianov *et al.* 2026 for a similar observation). In contrast, at high salinities, the sands are more conductive than the clay-rich and organic matter-rich samples. There is an intermediate pore water conductivity range ($\sigma_w \in [0.2, 1] \text{ S m}^{-1}$) for which the conductivity of all sediments are roughly equal. This points out the challenges of deriving porewater conductivity from electrical resistivity measurements alone.

4.2 Archie's law

For each sample, once we have the value of the formation factor and the porosity, we can determine the value of the porosity exponent (also called the cementation exponent) m of Archie's law $F = \phi^{-m}$ using $m = -\ln F / \ln \phi$ (F and m should be replaced by apparent values in presence of pyrite as suggested above). In Fig. 7, we plot the formation factor as a function of the porosity and we observe that we cannot fit a single Archie's law (i.e. a single value of m) to the complete data set.

It has been shown in the past that the cementation exponent m could increase with the CEC itself for clay-rich

media (see M.H. Waxman & L.J.M. Smits 1968; A. Revil *et al.* 1998) and the model developed above suggest that the apparent cementation exponent could decrease with the increase of the pyrite content (Fig. 8). Fig. 8 show the dependence of the m -exponent with the measured CEC of the sample. For sands and clays, the m -exponent increases monotonically with the measured CEC and the relationships have been parametrized. For peat and organic-matter rich soils and shallow sediments, we observe a very different relationship between the m -exponent and the measured CEC: a decrease of the apparent m -exponent versus the CEC. This decrease is likely explained through the presence of pyrite (Fig. 9). The framboidal pyrite (from the french word framboise, raspberry), typically present in coastal soils and sediments, are 5 to 20 μm in diameter (see Fig. 9 and H. Ohfuji & D. Rickard 2005, and see discussion above).

To confirm the decrease of the apparent cementation exponent with the presence of magnetite or pyrite, we performed an additional set of experiments. We took a sandy soil from our collection (porosity 0.35, CEC 3.7 meq per 100 g, grain size around 190 μm) and we mixed it with various volume fractions of a synthetic magnetite (in fine powders). The experiments are shown in Fig. 10. We observe, in qualitative agreement with our model, that the apparent cementation exponent decreases with the addition of magnetite and can reach values below 1.0 for a critical magnetite content of ~ 4 per cent. Using eqs (23) and (24),

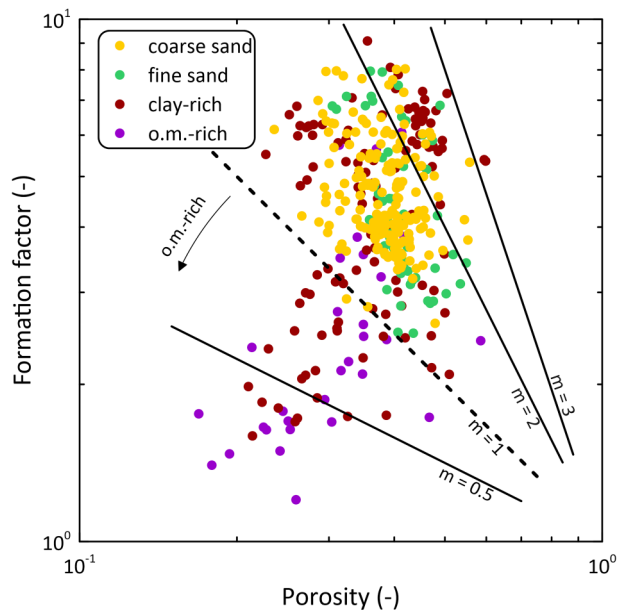


Figure 7. Archie's law between the intrinsic formation factor and the (connected) porosity, for the samples. The data set exhibits a broad variation in the cementation/porosity exponent m with low values for the organic matter-rich sediments to high values for part of the clay-rich sediments. At first, it is unexpected that the cementation/porosity exponent m can be smaller than one. An explanation of this behaviour is due to the presence of metallic particles (framboidal pyrite in the present case) as extensively discussed in the main text.

$m = 1.70$ [from the extrapolation done in Fig. 10(d), a porosity of 0.50 ± 0.04 for the mixture between the soil and the magnetite], we obtain a critical value of the magnetic content of 21 per cent, higher than the observed value of 4 per cent. The consistency between the model and the data is therefore only qualitative at this stage.

4.3 Normalized chargeability versus quadrature conductivity

In Fig. 11, we plot the (partial) normalized chargeability (M_n) versus the quadrature conductivity (σ'). They both characterize the polarization phenomenon *per se*. As predicted by eq. (12), they are linearly related to each other with a slope of 4.1 (versus 4.4 for the theory using $\alpha \approx (2/\pi) \ln A$ and A corresponding to three orders of magnitude). This small difference (nearly insignificant in a log-log plot) can be explained by the fact that the phase, strictly speaking, does not follow a Constant Phase Model (CPM). In the field, looking at the frequency dependence (dispersion) of the in-phase conductivity or looking at the quadrature component are equivalent in terms of information content as long as the level of noise on the two measurements are equivalent.

4.4 Normalized chargeability versus surface conductivity

According to eqs (6) and (8) (non-fractal model) or eqs (9) and (11) (fractal model), the normalized chargeability and surface conductivity are proportional to each other and both cases their ratio define the dimensionless number R . In Fig. 12, we plot the normalized chargeability (M_n) versus the surface conductivity

(σ_s). These parameters characterize polarization and conduction, respectively. They are associated with the properties of the electrical double layer and show a moderate correlation to each other. This is consistent with the prediction of the dynamic Stern layer model indicating a slope of $R = 0.10 \pm 0.02$. However, there is a substantial number of data points from all four lithological groups above this trend indicating the presence of an additional polarization source in the set of samples. This polarization source corresponds to the presence of metallic particles, for example, magnetite or pyrite. The role of pyrite is further discussed below.

4.5 Surface conductivity and normalized chargeability versus CEC

Next, we plot both the normalized chargeability (M_n , Fig. 13) and the surface conductivity (σ_s , Fig. 14) versus the measured CEC divided by the tortuosity of the pore space (obtained by the product of the formation factor and the porosity, see eqs 6 and 8 for the non-fractal model). We see a fair correlation between these electrical properties and the CEC of the sediment samples and the trends are roughly consistent with the classical dynamic layer Stern model developed in A. Revil *et al.* (2017), but the variation around this model is very large, especially for the organic matter-rich samples. The scatter for the OM-rich samples will be explained further below because of the presence of framboidal pyrite. We have no explanation for the scatter associated with the coarse sands.

In Fig. 15, we plot the quadrature conductivity and normalized chargeability versus the excess of charge per unit volume as prescribed by the fractal-based model of H. Wang & A. Revil (2020). We found a reasonable correlation with an exponent $m_s = 0.79$ for the interfacial Archie's law to the predicted value of 0.78 ± 0.10 by H. Wang & A. Revil (2020, their table 1). This yields to a mean value of the pore volume fractal dimension D_f equal to ~ 2.76 in agreement with the upper value observed for soils by C. Ahl & J. Niemeyer (1989) using the water retention curve.

4.6 Volume averaged DSL model versus fractal-based DSL model

By using both DSL models on the same large data set, it may be possible to evaluate whether one is better than the other. Fig. 15 seems to indicate that the fractal-based model seems to slightly better fit the data compared to the classical DSL model but as the difference is small a definitive answer could not be brought to this question yet. This is at least partly because not all sediments are necessarily fractal-based. Further comparison will be needed in future studies.

4.7 Can we determine the formation factor at a single salinity?

This question was raised by A. Weller *et al.* (2013) (based on experimental evidences) and A. Revil (2013) (using the classical DSL model). Knowing the conductivity of the pore water, we have two measurements (the conductivity and the normalized chargeability) for two unknowns, the formation factor and the surface conductivity. In theory, we can easily determine these two unknowns from eqs (6) and (8) on the condition that the

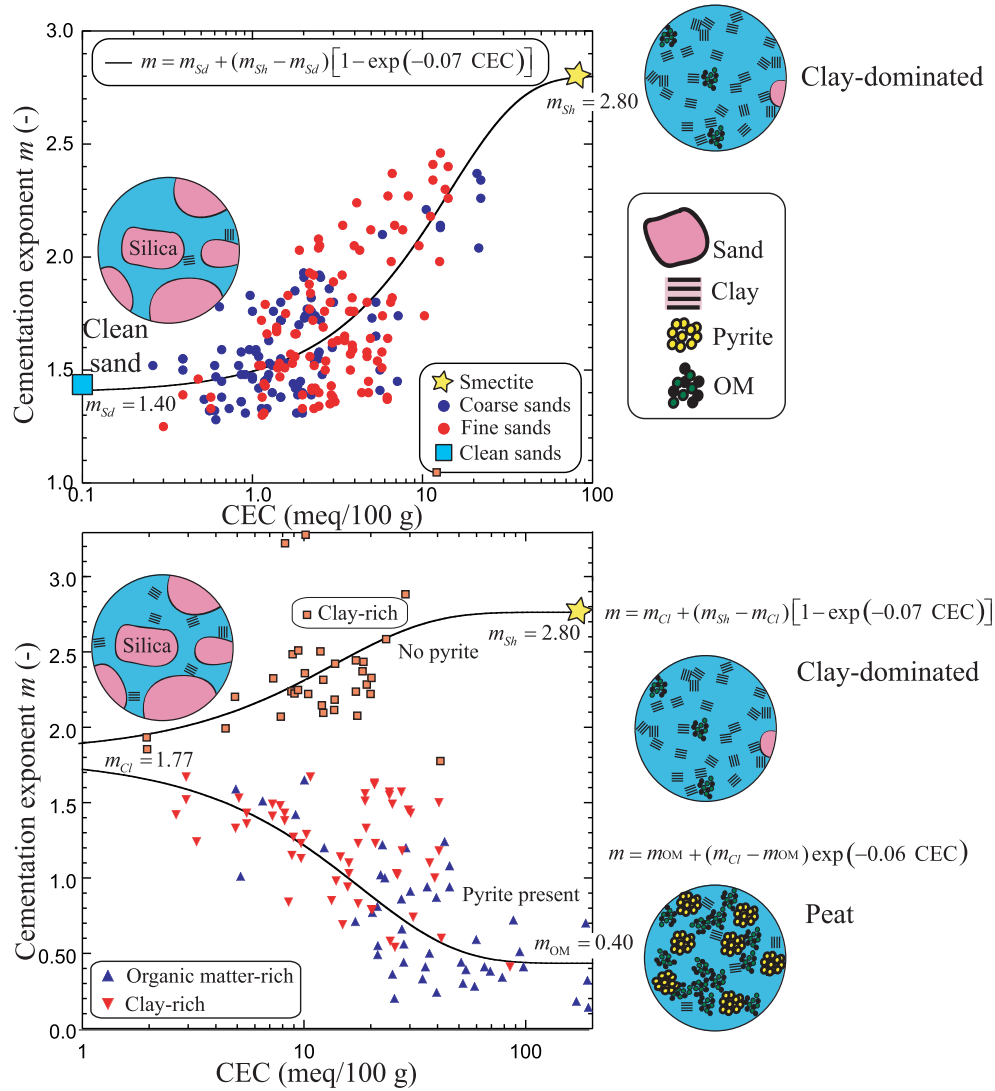


Figure 8. Deciphering the value of the cementation exponent in soils and shallow sediments. The star corresponds to the smectite end case with high CEC (typically 80 meq per 100 g) and cementation exponent close to 2.8 (A. Revil *et al.* 1998). As shown in the text, the presence of framboidal pyrite has a drastic effect on the value of the apparent cementation exponent, which can be smaller than unity. Note that the trend made by the clay-rich 'pyrite unlikely' data in the lower graph seems to align with the trend in the sand-rich upper graph. The quantities m_{sd} and m_{sh} are the two end-members first-Archie's exponents for the sand and shale end-members, respectively for the sand clay mixtures, m_{cl} and m_{om} are the two end-members first-Archie's exponents for the clay-rich and OM (Organic Matter)-rich end-members, respectively clayey and OM-rich facies.

salinity of the pore water is not too low for surface conductivity to completely dominate the response of the soil and sediments and assuming we can neglect the effect of metallic particles like pyrite. If this is the case, indeed, the conductivity is equal to the surface conductivity and the normalized chargeability is just the surface conductivity times $R = 0.10$ as illustrated in Fig. 12. Thus, for intermediate salinities, we can determine the formation factor and the surface conductivity very well. Fig. 16 shows the determination of the formation factor based only on the intermediate of the three salinities (horizontal axis) versus the formation factor traditionally determined from the plot of the conductivity of the sediment versus the conductivity of the pore water using the full salinity range (Fig. 6). The agreement is reasonably good. It indicates that the model provides a fair determination of the formation factor but that determination

remains far from perfect. The discrepancy may be found in some assumptions made in the model or the fact that some of the CEC may be located in dead-ends.

In Fig. 17, we plot the formation factor (determined from the full salinity range) versus the surface conductivity of the samples. Such a cross-plot allows to roughly distinguishing between the three lithology groups used in our study (fine and coarse sands, clay and OM-rich). It becomes apparent that the clay group consists of two rather distinct parts with a higher and lower formation factor. The same division appears to be present in the organic rich samples, albeit less clear. Both sandy groups (fine and coarse) appear to be more continuous and gradual in their distribution. Such an approach could be used with field data to do soil and shallow sediments classification using clustering techniques. The development of drone, towed or airborne

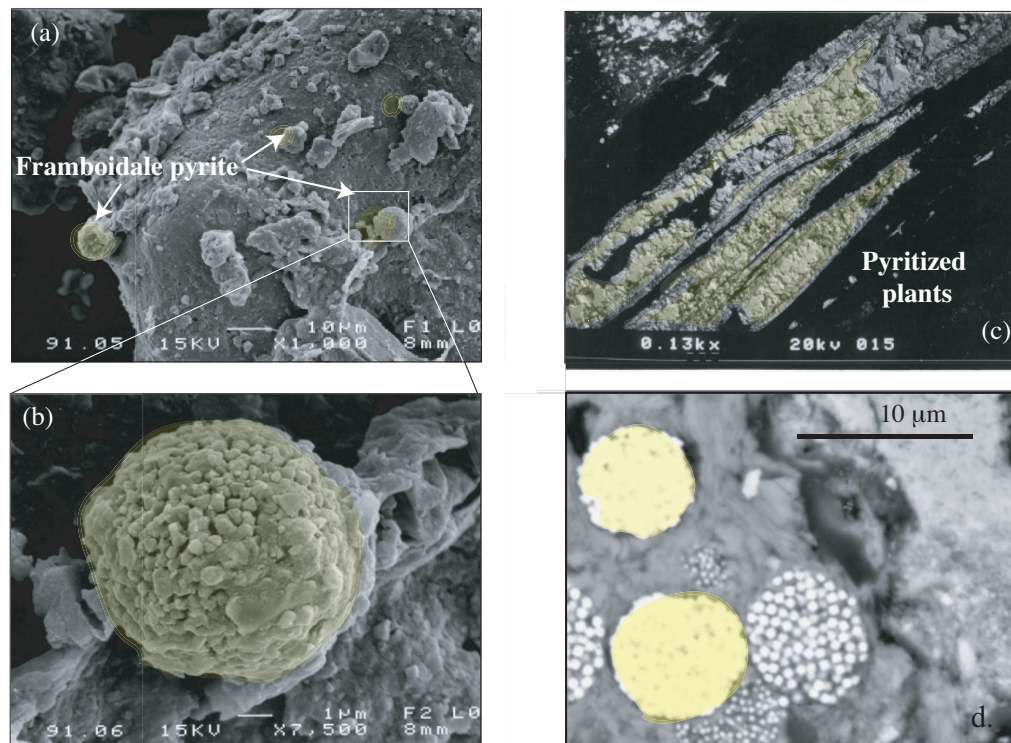


Figure 9. Presence of pyrite in sediments and soils of the Netherlands, taken from a Holocene fluvial channel deposit near Alphen aan den Rijn, western Netherlands. (a, b) Pictures showing the presence of framboidal pyrite at the surface of a sand grain. The size of the framboids is typically between 5 and 20 μm (modified from J. Griffioen *et al.* 2016). (c) For organic matter-rich (peat) samples, a relatively high pyrite content can generally also be expected, occurring as pyritized plant remains (modified from the review paper on this topic by J. Griffioen *et al.* 2016). (d) Picture of the framboidal pyrite.

Time Domain ElectroMagnetic (TEM) surveying including the effect of induced polarization in the interpretation of the data would foster the application of induced polarization to agriculture for the determination of soil quality.

4.8 What is the role of the organic matter?

The CEC of some organic matter-rich samples may reach nearly 100 to 200 meq per 100 g and is therefore very high compared to reference values of most clay minerals (80 meq per 100 g at the highest). Most samples fall in the range of 50 to 90 meq per 100 g. From the results presented previously (e.g. Figs 13, 14 and 15), its influence on both surface conductivity and normalized conductivity/quadrature conductivity is clear, causing higher conduction and polarization with respect to samples with low organic matter content.

Finally, we analysed the relationship between the TOC and the CEC for 33 samples from the Dutch subsurface that are rich in organic matter. The results are shown in Fig. 18. As expected, we see a linear trend between two poles corresponding to the organic matter pole (CEC = 160 meq per 100 g) and a clay pole (~ 8 meq per 100 g). The CEC of the organic matter is higher than the CEC of smectite (the clay mineral characterized by the highest CEC values, i.e. 80–120 meq per 100 g) and therefore should be accounted for in CEC calculations.

4.9 Time-domain versus frequency-domain normalized chargeability

Usually, in field conditions, induced polarization is measured in the time-domain. The electrical current is injected for a period T (between two electrodes A and B), and the secondary voltages recorded through two other electrodes M and N. The voltage decay is integrated over a time-window Δt after a dead-time t_D . It follows that the normalized chargeability obtained in the field $M_n(\text{TD})$ is necessarily smaller than the normalized chargeability defined in the frequency-domain as $M_n(\text{TD}) \leq M_n(\infty, 0) \equiv \sigma_\infty - \sigma_0$ (see L. Piolat *et al.* 2025, and a complete analysis in Appendix A). We consider here an additional set of experiments to compare the two quantities to determine the amplification factor $a \equiv M_n(\infty, 0)/M_n(\text{TD}; T, \Delta t, t_D)$. In order to be meaningful, we used 46 samples in our collection and we used the following values of the time-domain parameters $T=1\text{s}$, $\Delta t = 0.10\text{s}$, $t_D = 0.15\text{s}$. The time-domain data were recorded with an IRIS-TERRA, an input current of 1 mA and the same sample holder as before. The results are shown in Fig. 19 and yields $a = 10 \pm 4$. The scattering in Fig. 19 is because the spectra may be very different between sandy and clayey soils and sediments. That said, our results imply that a substantial correction of the field data is required to bring them back to the laboratory-derived frequency-domain parameters.

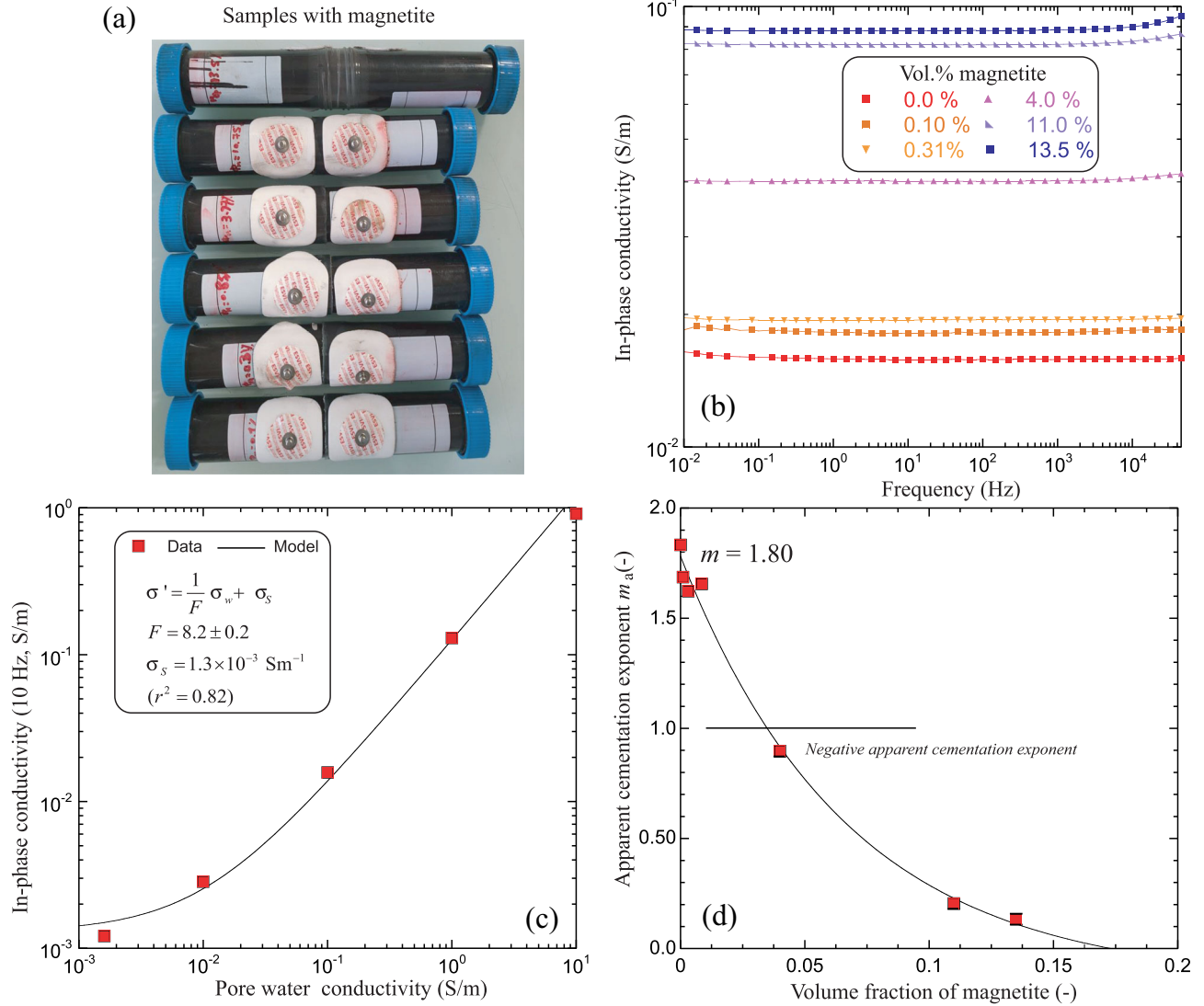


Figure 10. 12 additional experiments performed with a sandy soil (CEC = 3.7 meq per 100 g, porosity 0.35) mixed with various amounts of synthetic magnetite (from 0 per cent to 13.5 per cent in volume fraction). (a) Picture of the samples mixed with magnetite used for the experiments. (b) In-phase conductivity spectra at a pore water conductivity of 0.1 S m^{-1} (25°C , NaCl). (c) Conductivity of the sandy soil with the pore water conductivity at different salinities. These experiments are used to determine the formation factor and surface conductivity at 25°C in absence of magnetite. (d) Apparent cementation exponent m_a determined from the apparent formation factor versus the volume fraction of magnetite.

4.10 Specific effect of pyrite

In order to provide a last test to our classification in terms of pyrite presence, we plot in Fig. 20, the apparent cementation exponent of the samples characterized by a high likelihood of the presence of pyrite. As expected most of these samples have an apparent cementation exponent below 1.0. This confirms that the classification we used for the presence of pyrite is consistent with the model predictions.

5 FIELD APPLICATION: HOW TO MAP THE CEC FROM INDUCED POLARIZATION

In this section, we show how to apply the petrophysical relationships to interpret field measurements. We use the field data from a survey performed in March 2024 on Basse-Terre Island in the Guadeloupe archipelago (French West Indies, Lesser

Antilles, France, Fig. 21). The soils present at our test site are nitisols. They are characterized by iron-rich kaolinite/metalloxydites 1:1 clay assemblages with a clay content > 30 per cent and a CEC typically around 20 meq per 100 g and with generally low organic matter content (<2 per cent) and pyrite (J. Sansoulet *et al.* 2007; J. Sierra & L. Desfontaines 2018).

We used an ABEM-SAS-1000 resistivity meter with cables with a spacing of 50 cm between the stainless steel electrodes (5 mm in diameter and 120 mm in length, 10 cm immersed into the ground). The injected current was 100 mA. To ensure a good contact between the electrodes and the ground, clay and saline water were added to reduce the contact resistance to be less than $1 \text{ k}\Omega$ when needed. To improve the signal-to-noise ratio for chargeability measurements,

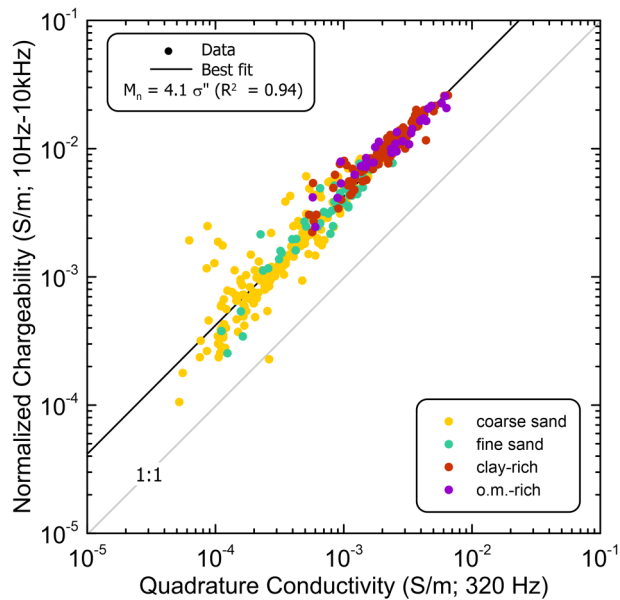


Figure 11. Normalized chargeability versus quadrature conductivity. The black plain line corresponds to the best fit of the data. The fitted slope 4.1 ± 0.5 is close to the theoretical value obtained from the main text ($\alpha = 4.4$).

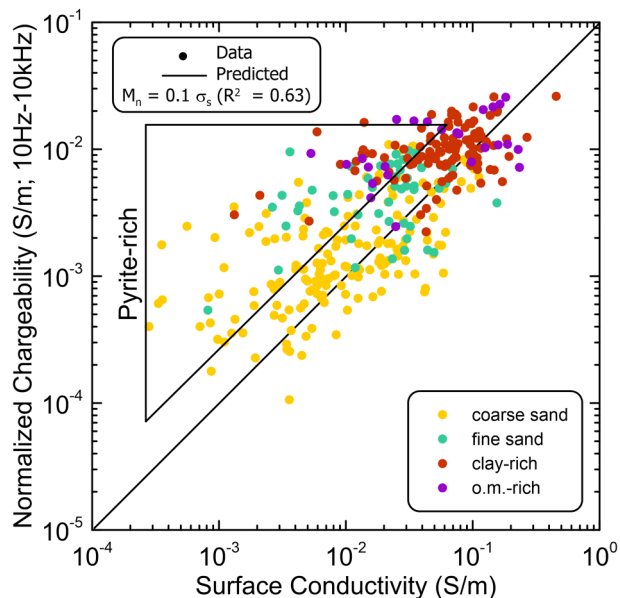


Figure 12. Normalized chargeability versus surface conductivity. The straight solid line corresponds to the prediction of the dynamic Stern layer model such as presented by A. Revil *et al.* (2017), the coefficient of proportionality being the dimensionless number $R = 0.10$. The data above this line is likely affected by the presence of metallic particles (e.g. framboidal pyrite) present in the samples.

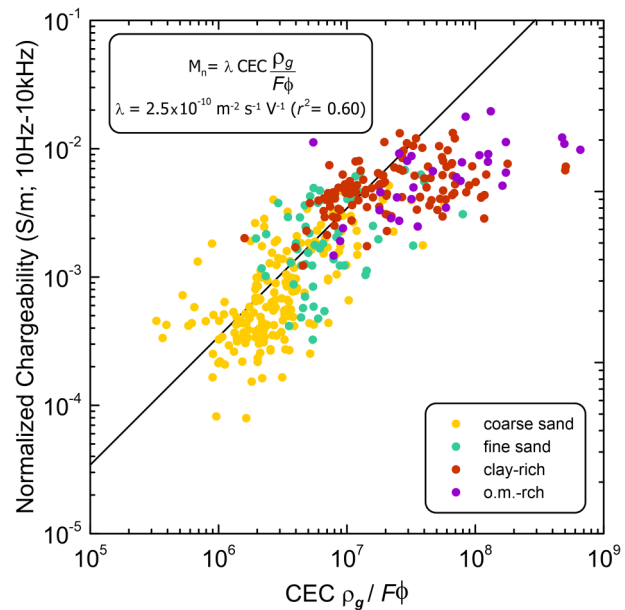


Figure 13. Normalized chargeability versus the product of the grain density times the CEC normalized by the bulk pore space tortuosity (a bulk grain density of 800 kg m^{-3} is used for the organic matter-rich samples). The tortuosity of the pore space is given by the product of the formation factor by the connected porosity. The black plain line corresponds to the prediction of the dynamic Stern layer model such as developed by A. Revil *et al.* (2017).

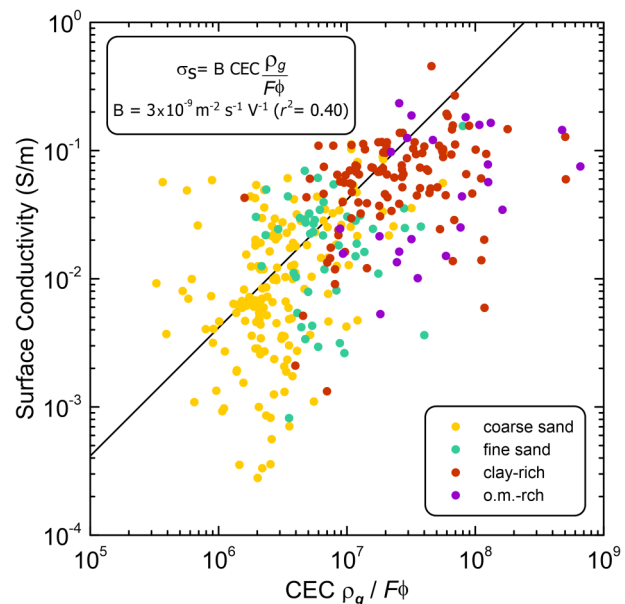


Figure 14. Surface conductivity versus the product of the grain density times the CEC normalized by the bulk pore space tortuosity (a bulk grain density of 800 kg m^{-3} is used for the organic matter-rich samples). The tortuosity of the pore space is given by the product of the formation factor by the connected porosity. The black plain line corresponds to the prediction of the dynamic Stern layer model such as developed by A. Revil *et al.* (2017).

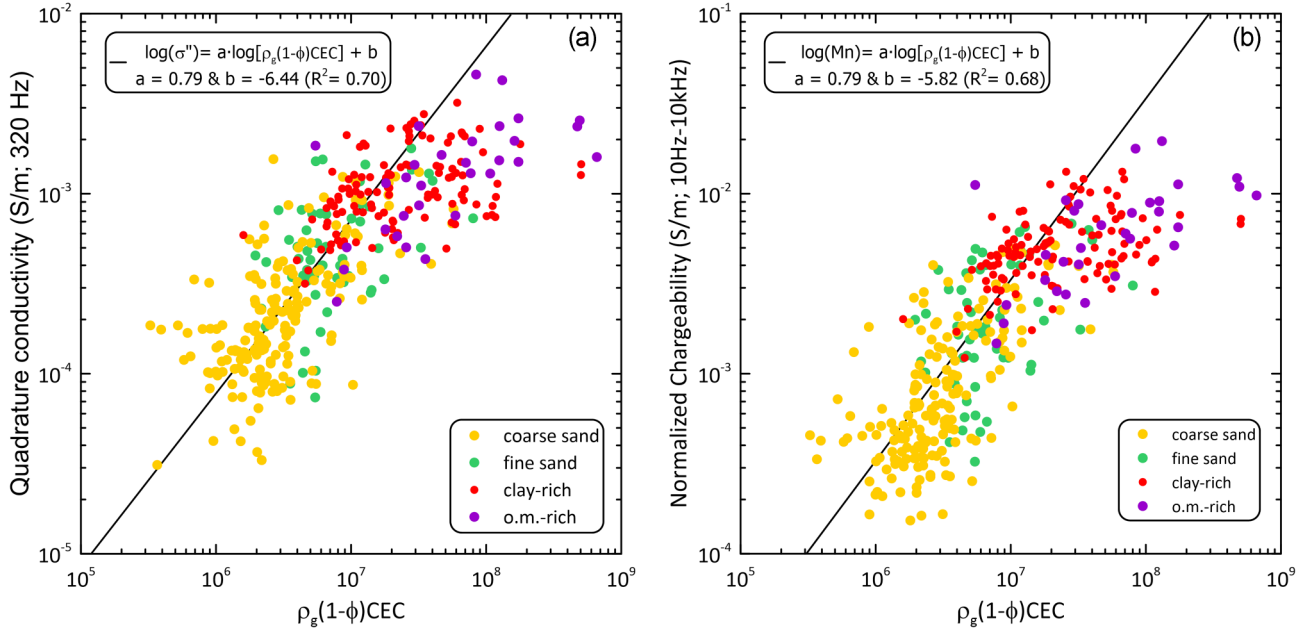


Figure 15. Quadrature conductivity and normalized chargeability *versus* the excess charge per unit total volume of porous material according to the fractal-based model of H. Wang & A. Revil (2020). (a) Quadrature conductivity *versus* the charge per unit volume of soil (a bulk grain density of 800 kg m^{-3} is used for the organic matter-rich samples). (b) Normalized chargeability *versus* the charge per unit volume of porous material (a bulk grain density of 800 kg m^{-3} is used for the organic matter-rich samples). The black plain line corresponds to the prediction of the extension of the fractal model developed by H. Wang & A. Revil (2020) (see S. Qiang *et al.* 2025, for details). The exponent of the associated interfacial Archie's law is equal to $m_s = 0.79$, close to the values provided in H. Wang & A. Revil (2020).

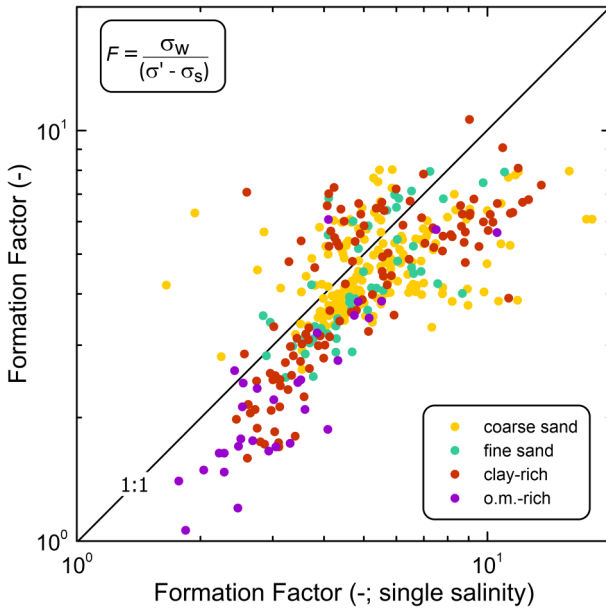


Figure 16. Determination of the formation factor using the full salinity range (three salinities) *versus* the formation factor determined at a single (intermediate) salinity using the relationship between the normalized chargeability and the surface conductivity.

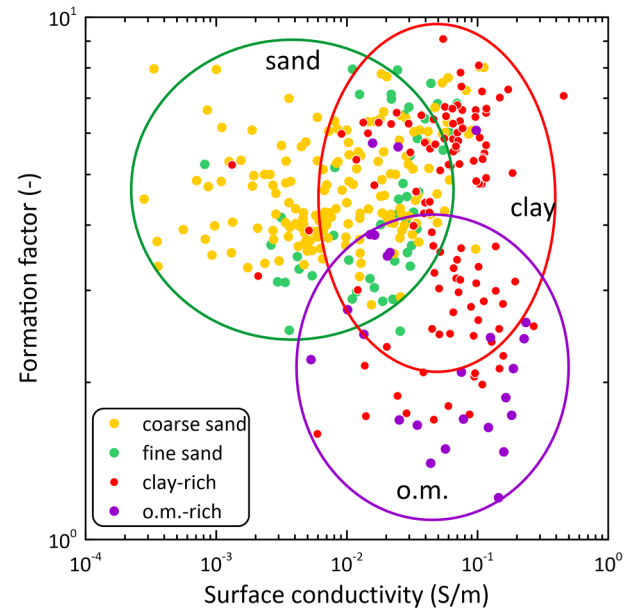


Figure 17. Cross-plot of the formation factor (determined from the full salinity range) *versus* the surface conductivity. This type of cross-plot can be used to differentiate the facies since the three (sandy, clay-rich and organic matter (om)-rich) poles can be separated using such a cross-plot. This approach could be applied to field data using clustering techniques including the position of the data points in the tomograms.

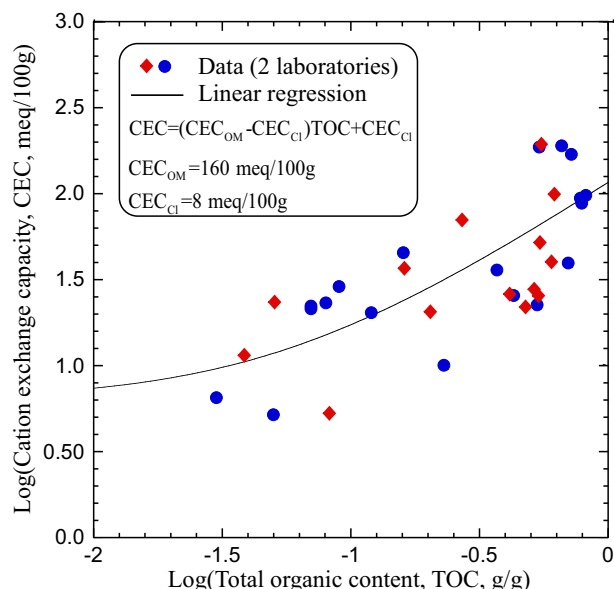


Figure 18. Relationship between the CEC and the TOC for 33 organic-rich samples. A linear trend (nonlinear in a log–log scale) is used to determine the CEC of the pure organic matter pole ($CEC_{OM} = 160$ meq per 100 g) and the CEC of the clay fraction associated with the organic matter ($CEC_{Cl} = 8$ meq per 100 g). The two data sets have been obtained in two distinct laboratories, EDYTEM in France for the circles (19 samples, dried at 58°C), and ISOLAB B.V. in the Netherlands for the diamonds (14 samples, dried at 100°C). Tests on the same samples indicate that the EDYTEM data may need a correction factor by a factor 0.51. The application of this correction factor (not used here) does not change the conclusions.

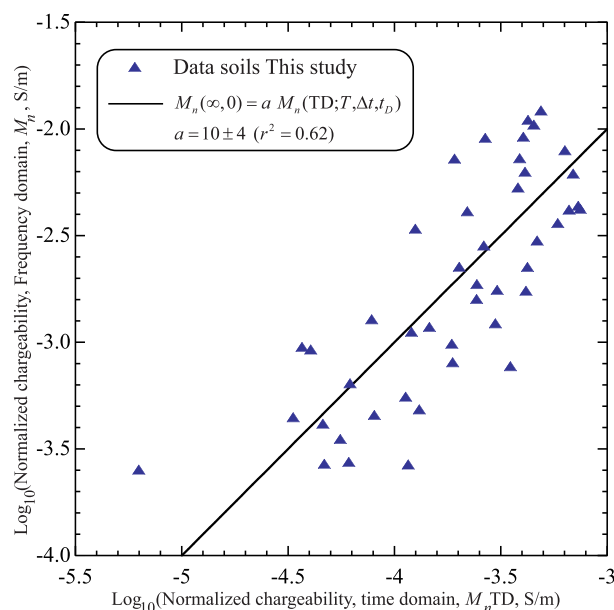


Figure 19. Comparison between the normalized chargeability measured in the frequency-domain (1 Hz–10 kHz) and the normalized chargeability obtained in the Time-Domain (TD) with the following set of parameters: Current injection 1 s, dead time 150 ms, duration of the time window 100 ms). The experiments were done at a pore water conductivity of 0.9 S m^{-1} at 25°C .

non-sorted Wenner protocol was used to avoid electrode polarization issues. Indeed, it may take up to five minutes for a current electrode to discharge and therefore during this time, it cannot be used as a voltage electrode. The protocol used 472 measurements for 64 electrodes (long profiles) and 139 measurements for 32 electrodes for the short profiles parallel to each other. The survey comprised 166 profiles of 32 electrodes and 7 long profiles using 64 electrodes and roll-along of the electrodes (Fig. 21).

The acquired profiles underwent filtering on the apparent resistivity pseudo-sections before 2.5-D inversion processing, which was performed up to the third iteration using the least-squares method. The full data set comprised 33 545 apparent electrical conductivity and time-domain chargeability data. The 2.5-D inversion of apparent resistivity and chargeability was performed using the Res2Dinv software, and 20 cm depth iso-maps were extracted from 3-D linear interpolation between the 2.5-D profiles. Only the data from uppermost depth iso-map (upper 20 cm) corresponding to the organic-rich soil are shown and used below. Three samples were taken from the field and analysed for their CEC using the same technique as used for the sediment samples from the Netherlands (see position in Fig. 21 and physical properties in Table 2). We used only three samples since the soil in the field was found to be rather homogeneous (banana plantation).

The inverted conductivity and normalized chargeability maps are shown in Fig. 22. Since the current is injected for a short period of time ($T_{on} = 0.4 \text{ s}$), a correction on the inverted normalized chargeabilities should be performed to obtain the normalized chargeability for an infinite time of current injection (i.e. covering the full spectrum of frequencies). The connection between time-domain and frequency-domain induced polarization measurements was recently discussed in several papers (e.g. T. Martin *et al.* 2021). In the present case, the correction factor, a , we used is around 10 as derived from and shown in Figs. 19 and 23. The rationale for this amplification factor a is extensively discussed in Appendix A. It is an important point that was, for instance, overlooked in the recent study of P. McLachlan *et al.* (2024). P. McLachlan *et al.* (2024) used a three decade range in their analysis. If we consider that a soil is fractal between 0.1 and $100 \mu\text{m}$, this is four decades in grain sizes and therefore eight decades in frequencies (accounting for the relaxation time depending on the square of the grain size). Three decades range is therefore not enough and a correction factor should be used to go from three decades to eight decades (see discussion in S. Qiang *et al.* 2025, using the CPA model). The fact that we use a short injection time for the injected current explains why the apparent chargeabilities are small in our case.

The conductivity of the shallow groundwater is $460 \mu\text{S cm}^{-1}$ at 25°C (0.046 S m^{-1}), which is close to the conductivity of the water in equilibrium with the three samples collected at the site (see Fig. 22 and Table 2) and determined to be 0.048 S m^{-1} (25°C). The soils of Basse-Terre Island (Guadeloupe) are usually clay-dominated (J. Sierra & L. Desfontaines 2018). In the study area, they overlie a volcanic rock substratum consisting of andesites from the GM_3 group (see G. Boudon *et al.* 1988). They are relatively poor in organic matter and rich in kaolinite. Kaolinite is characterized by a low CEC ($\sim 10\text{--}20$ meq per 100 g). This mineral is expected to control the CEC of these soils.

In order to determine the water content and the CEC from the geophysical data, we use a pore water conductivity of 0.048 S

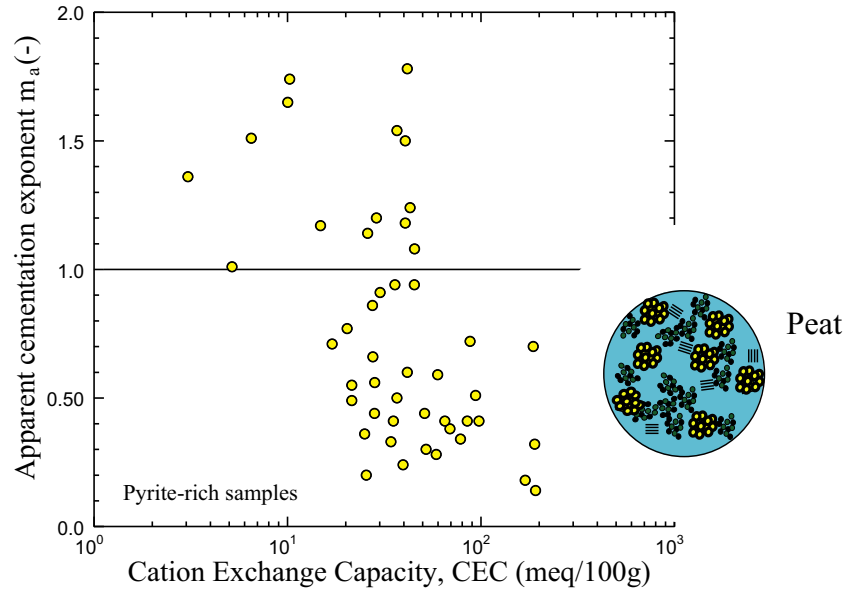


Figure 20. Apparent cementation exponent versus CEC for the samples where it was assigned that the presence of pyrite was likely or very likely. We observe that most of these samples have an apparent cementation exponent below 1.0 which confirms the modelling



Figure 21. Presentation of the test site. (a) Test site. The lines correspond to the induced polarization profiles made in the survey. The three circles correspond to the position of the three soil samples collected in the field. (b) Position of the field data in Guadeloupe Island. (c) The site is characterized by the presence of some buried boulders, some of which are of archaeological significance (pre-Columbian rock engravings). Nowadays, the investigated area is used mostly for Creole garden-type subsistence farming, with extensive banana growing areas.

Table 2. Characteristics of the three samples (halloysite-rich nitisols) used in the field survey, see sampling position in Fig. 20. The pore water conductivity in equilibrium with the samples is 0.0478 S m^{-1} (25°C). The in-phase conductivity and normalized chargeabilities are given at this pore water conductivity. The formation factors are determined at a high salinity corresponding to a pore water conductivity of 14.3 S m^{-1} (NaCl , 25°C).

Sample ID	ϕ (-)	CEC (meq per 100 g)	F (-)	σ' (1 Hz, S m^{-1})	$M_n(1-100 \text{ Hz}) (\text{S m}^{-1})$	$M_n(0.1 \text{ Hz}-10 \text{ kHz}) (\text{S m}^{-1})$
S01	0.50	15.9	4.6	0.0615	3.7×10^{-3}	10×10^{-3}
S02	0.53	16.2	4.0	0.0411	2.2×10^{-3}	5.5×10^{-3}
S03	0.66	20.6	5.2	0.0401	7.9×10^{-3}	3.7×10^{-3}

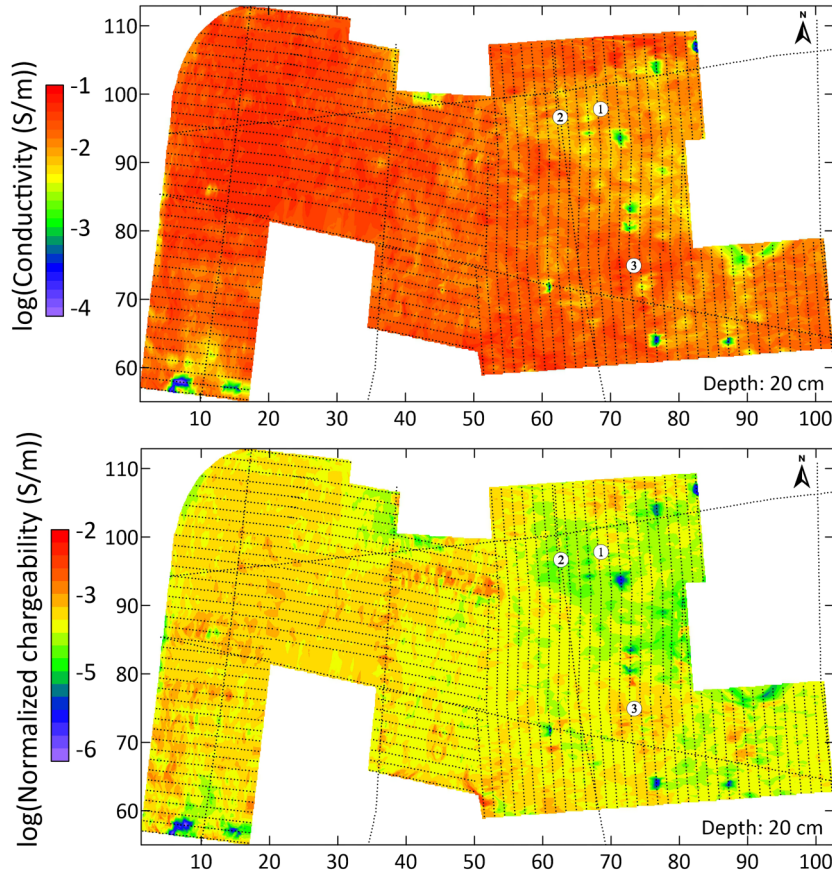


Figure 22. Map of the electrical conductivity and normalized chargeability at a depth of 20 cm using time-domain induced polarization with a current injection window of $T_{on} = 0.4$ s. The spots characterized by low values of the conductivity and normalized chargeability correspond to the presence of boulders. The position of the three samples is also indicated. There are 9 106 data points at a depth of 20 cm. After performing linear interpolation between these profiles in COMSOL, the data set is expanded to 268 137 points.

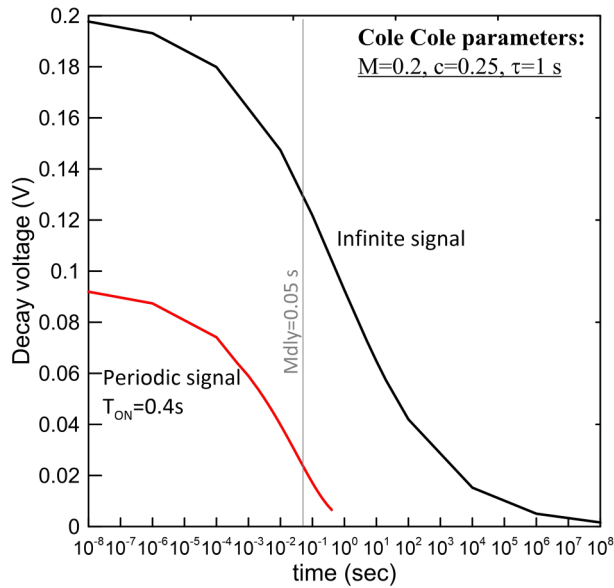


Figure 23. Decay voltage curves for two values of the current injection period T_{on} (comprised between 0.4 s and an infinite time) given a Cole-Cole complex conductivity model. This figure shows the necessity to correct the field data by an amplification factor ‘a’.

m^{-1} (25°C). We use the volume-averaged-based DSL model by replacing $\phi^m s_w^n$ by θ^m assuming *ipso facto* that $m = n$ (an assumption broadly done in interpreting downhole measurements in the realm of the oil industry, see also A. Revil *et al.* 2023a, b, 2024; J. El Alam *et al.* 2024, 2025). In this case, the classical DSL model (the same could be done with the fractal-based model) can be written as,

$$\sigma_{\infty} = \theta^m \sigma_w + \theta^{m-1} \rho_g B \text{CEC}, \quad (27)$$

$$M_n = \theta^{m-1} \rho_g \lambda \text{CEC}. \quad (28)$$

Eqs (26) and (27) can be inverted to get the water content and the CEC from the conductivity and the normalized chargeability knowing the conductivity of the pore water σ_w . The water content and the CEC maps are shown in Figs 24 and 25 depending on the values of the correction factor (1 or 10), respectively. Some areas are characterized by specifically low values of conductivity and normalized chargeability corresponding to the presence of boulders or boulder accumulations embedded in the soil. Fig. 26 shows a cross-plot between the normalized chargeability and the conductivity for the data at a depth of 20 cm, which corresponds to the depth of soil sampling. The trend (much below the slope corresponding to $R = 0.10$) indicates that surface conductivity is not dominant and therefore that we are at an intermediate salinity, a condition required to get good estimates of the water

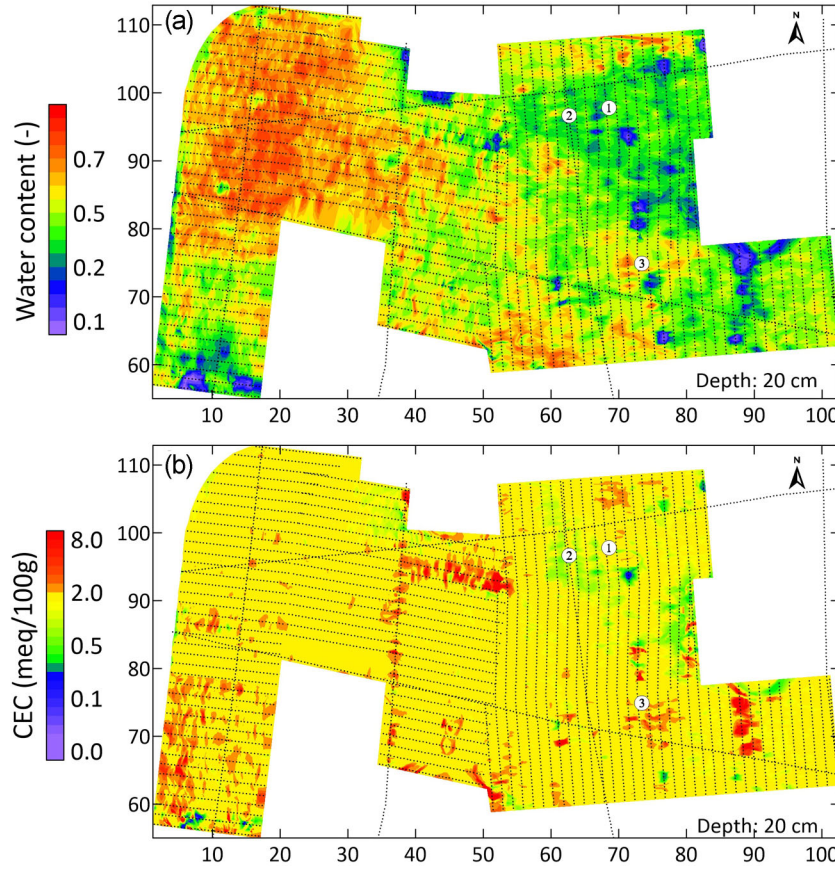


Figure 24. Map of the water content θ (a) and CEC (b). The spots characterized by low values of the water content (in dark) correspond to the presence of boulders (see Fig. 21). The position of the three samples is also indicated by the circles.

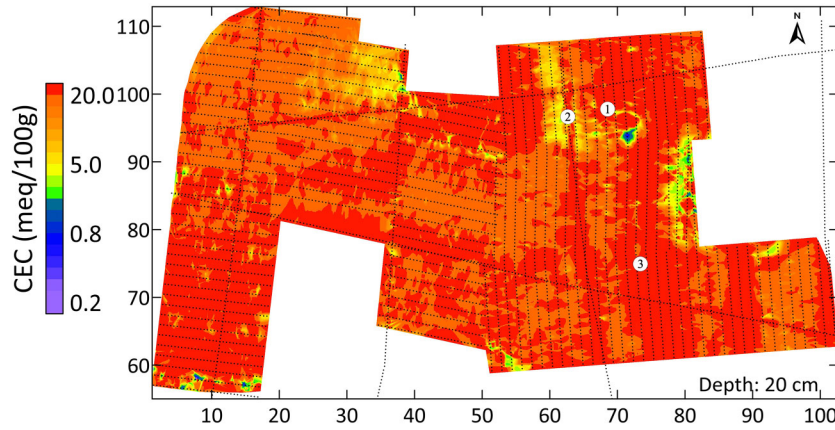


Figure 25. Map of the CEC using a correction factor of 10 for the normalized chargeability.

content and CEC values as discussed above. Indeed, if surface conductivity dominates the overall conductivity response, the conductivity and normalized chargeability are just proportional to each other and the two methods cannot be used to assess the formation factor or θ^m in unsaturated conditions. Note that the value of R is unchanged since it is the ratio of the two mobilities λ and B (i.e., $R = \lambda/B$). The ratio between the normalized chargeability and the conductivity is not the ratio between the normalized chargeability and the surface conductivity.

Finally, in Fig. 27, we show the complex conductivity spectra of three core samples collected in the field using a four

electrodes approach as shown in the insert of the figure. The impedance meter used for the measurement remains the same as before. These spectra have, if compared to the clay rich examples from the Netherlands shown in Fig. 5(b), a lower in-phase conductivity, being a closer match to the fine sand lithology group shown in Fig. 5(a). The in-phase and quadrature conductivity data of the Guadeloupe samples are consistent with the field survey. The CEC determined from the field survey is also consistent with the CEC of the three soil samples, which is in the narrow range 16 to 21 meq per 100 g (see Table 2). This range is consistent with those independently

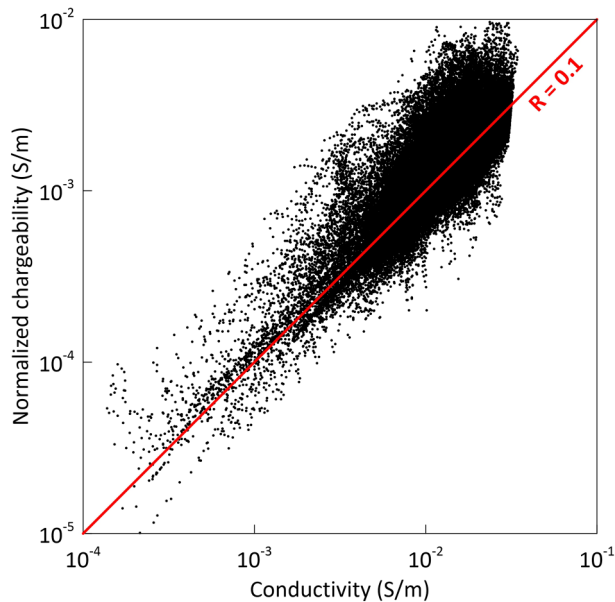


Figure 26. Correspondence between the normalized chargeability and the electrical conductivity from the field data. The trend is localized along the trend characterized by $R = 0.10$ meaning that the conductivity is dominated by the surface conductivity.

reported by J. Sansoulet *et al.* (2007) and J. Sierra & L. Desfontaines (2018).

Finally, in Fig. 27, we show the complex conductivity spectra of three core samples collected in the field. These spectra have, if compared to the clay rich examples from the Netherlands shown in Fig. 5(b), a lower in-Phase conductivity, being a closer match to the fine sand lithology group shown in Fig. 5(a). The in-phase

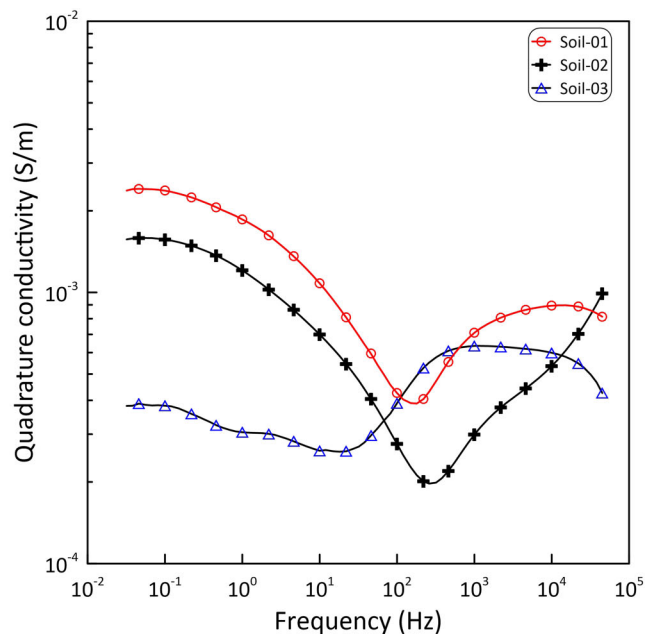
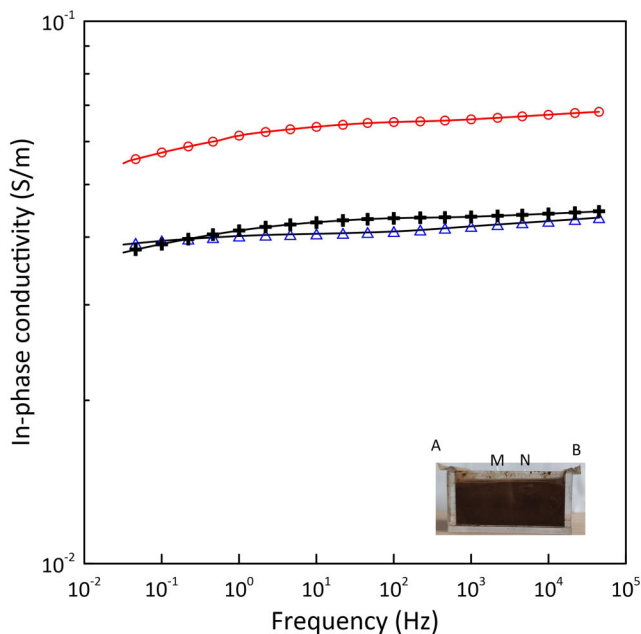


Figure 27. Spectral induced polarization data of the three core samples used in our analysis (see Table 2 for a summary of their properties). In principle, the effect of the electrode polarization can be neglected since we use a four-electrodes approach separating the current and voltage electrodes. Furthermore, the effect of the measuring setup has been removed from the measured response.

and quadrature conductivity data of the Guadeloupe samples are consistent with the field survey. The CEC determined from the field survey is also consistent with the CEC of the three soil samples, which is in the narrow range 16 to 21 meq per 100 g (see Table 2). This range is consistent with those independently reported by J. Sansoulet *et al.* (2007) and J. Sierra & L. Desfontaines (2018).

6 CONCLUSIONS

Previous works dealing with understanding induced polarization of soils and shallow sediments have resulted in empirical relationships based on linear correlations that are site specific. In this paper, we have developed universal relationships that are based on the underlying physics and that can be tested against a large data set. Two DSL models have been considered, one based on the volume averaging method and the second based on fractal theory. In order to test these relationships from these two models, we have performed spectral induced polarization measurements over 400 samples from unconsolidated sediment from the Netherlands, covering a broad range of soils and shallow sediment types. In addition, three (clay-rich) nitisols from Basse-Terre Island, Guadeloupe, were sampled. For simplicity, we grouped the Dutch samples in four categories: fine sands, coarse sands, clay-rich samples and organic-matter-rich soils. The measurements were performed at 25°C and at three salinities with NaCl solutions. The comparison between the data and the DSL conduction/polarization models leads to the following conclusions:

- (1) The organic matter is characterized by high CEC values (160 meq per 100 g for the end-member), low particle density and high conductivity and polarization at low salinities.

(2) The predictions of the DSL models explain the dependence between the normalized chargeability (characterizing polarization) and the surface conductivity (characterizing conduction in the double layer) fairly well. It explains also the dependence of the surface conductivity, normalized chargeability and quadrature conductivity with the CEC of the soils and shallow sediments. The constant-phase model (expected for the fractal case) provides also a relationship between the normalized chargeability and the quadrature conductivity that is in line with our new observations.

(3) The presence of framboidal pyrite complicates the direct interpretation of the spectral induced polarization data.

(4) The fractal based model seems to overperform with respect to the volume-averaged based DSL model regarding the relationship between the normalized chargeability and surface conductivity versus the CEC but further tests are required before to get a definitive answer.

A field application is performed to demonstrate how time-domain induced polarization can be used to assess, at the field scale, the CEC of soils and shallow unconsolidated sediments. It requires however the knowledge of the pore water conductivity or a calibration over multiple points. An important point is addressed when comparing the normalized chargeability obtained in time-domain induced polarization (with a current injection duration T) with the normalized chargeability obtained with frequency-induced polarization. We demonstrate that these are not equal. Furthermore, we provide a method to determine the amplification factor used to correct for the field data in time-domain induced polarization. We obtained realistic values of the field CEC values that are consistent with independent laboratory data on three soil samples (nitisol-type, CEC around 15–20 meq per 100 g).

ACKNOWLEDGMENTS

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DATA AVAILABILITY

The data used in this paper are available upon request to the corresponding author (A. Revil).

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APPENDIX A. FREQUENCY- AND TIME-DOMAIN NORMALIZED CHARGEABILITIES

As explained in the main text, the normalized chargeability determined from the time-domain measurements is different from the normalized chargeability determined using the complex conductivity spectra. In this appendix, we further explain this difference using a complex conductivity Cole-Cole model written as,

$$\sigma^* = \sigma_\infty - \frac{M_n(\infty, 0)}{(1 + i\omega\tau)^c}, \quad (\text{A1})$$

$$M_n(\text{FD}) \equiv M_n(\infty, 0) = \sigma_\infty - \sigma_0. \quad (\text{A2})$$

The time-domain response for an infinite charging time (current injected in the time window $[-\infty, 0]$) is obtained using the Mittag-Leffler function,

$$M(t) = \frac{M_n}{\sigma_\infty} \sum_{k=0}^{\infty} \frac{(-1)^k \left(\frac{t}{(1-m)^{1/c}\tau} \right)^{kc}}{\Gamma(1 + kc)}. \quad (\text{A3})$$

However in our laboratory time-domain experiments and field survey, the current is injected over a period T (i.e. in the time interval $[-T, 0]$). Since the system is linear and causal, we can use the superposition principle and the time-domain response is given by,

$$M_{\text{TD}}(t, T) = M(t) - M(t - T). \quad (\text{A4})$$

Furthermore, the normalized chargeability is obtained by integrating the decay (normalized) voltage over a time window following the dead time interval $[0, t_1]$, that is,

$$M_n(\text{TD}) = \sigma_\infty \int_{t_1}^{t_2} M_{\text{TD}}(t, T) dt, \quad (\text{A5})$$

with $(t_1, t_2) \ll \tau$. We use $t_2 - t_1 = 0.1\text{s}$. The amplification factor a is defined as,

$$a \equiv \frac{M_n(\text{FD})}{M_n(\text{TD})} \geq 1. \quad (\text{A6})$$