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Key Points:

- Geophysical methods offer new opportunities for soil structure characterization at larger spatial and temporal scales
- Combinations of geophysical methods expand the range of quantifiable soil structural attributes from pores to mechanical properties
- Geophysical methods can capitalize on surrogate dynamic hydrological variables as integrative indicators for soil structure at field scales

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A Review of Geophysical Methods for Soil Structure Characterization

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Abstract The growing interest in the maintenance of favorable soil structure is largely motivated by its central role in plant growth, soil ecological functioning, and impacts on surface water and energy fluxes. Soil structure pertains to the spatial arrangement of voids and solid constituents, their aggregation, and mechanical state. As a fragile product of soil biological activity that includes invisible ingredients (mechanical and ecological states), soil structure is difficult to define rigorously, and measurements of relevant metrics often rely on core samples or on episodic point measurements. The presence of soil structure has not yet been explicitly incorporated in climate and Earth systems models, partially due to incomplete methodological means to characterize it at relevant scales and to parameterize it in spatially extensive models. We seek to review the potential of harnessing geophysical methods to fill the scale gap in characterization of soil structure directly (via impact of soil pores, transport, and mechanical properties on geophysical signals) or indirectly by measurement of surrogate variables (wetness and rates of drainage). We review basic aspects of soil structure and challenges of characterization across spatial and temporal scales and how geophysical methods could be used for the task. Additionally, we propose the use of geophysical models, inversion techniques, and combination of geophysical methods for extracting soil structure information at previously unexplored spatial and temporal scales.

1. Introduction

Soil structure is defined as the spatial arrangement of solid constituents (minerals and organic matter) and voids of soil (Dexter, 1988) and is a reflection of biological activity (earthworms and roots), abiotic factors (freezing-thawing and wetting-drying), or results from tillage operations in the soil. Soil structure dynamics occurs at vastly different temporal scales (seconds to centuries), and although it fundamentally occurs at the pore scale, larger spatial scales (field, catchment, and region) become relevant since a wide range of hydrological and ecological soil functions are governed by soil structure (Stewart et al., 1990). A growing awareness of the key role that soil structure plays in providing soil ecosystem functions and services for all terrestrial surfaces (Bronick & Lal, 2005; Hamza & Anderson, 2005; Keesstra et al., 2012; Kibblewhite et al., 2008; Nawaz et al., 2013; Oertel et al., 2016; Zhang et al., 2007) has motivated recent attempts to quantify this important but elusive soil trait at relevant spatial and temporal scales (Besson et al., 2013; Guimarães et al., 2017; Keller et al., 2017). For most of these services and functions, a desirable soil structure is one that is able to support a wide range of biological activity ranging from microbial communities (Curtis et al., 2002) to mesofauna (e.g., earthworms and termites) and vegetation (Oades, 1993); organisms whose activity, in turn, contributes to the further development and maintenance of such desirable soil structure (Colombi et al., 2018; Young et al., 1998). This kind of well-developed soil structure facilitates fluxes of water and oxygen through the soil and makes these fluids available to plants, thereby, helping plant growth, promoting nutrient recycling and recharge of groundwater (Beven & Germann, 2013; Maximilian et al., 2009; Rabot et al., 2018). In contrast, a poor soil structure is one that restricts water infiltration and gas exchange, thereby, resulting in water runoff, soil erosion, and unfavorable anoxic conditions that limit plant growth and may trigger greenhouse gas emissions by anaerobic bacterial respiration (Berisso et al., 2012; Chen et al., 2014; Jordanova et al., 2011; Nawaz et al., 2013).

Several processes and mechanisms affect soil structure over multiple spatial and temporal scales. These processes are generally well known, including a reasonable level of quantitative understanding regarding soil structure degradation (e.g., compaction); however, our knowledge of soil structure generation, formation, and recovery remains limited (Keller et al., 2017). Mechanical and hydraulic stresses acting on or within the soil may generate or degrade soil structure. These stresses could either be produced by natural processes (both biotic and abiotic) or by human activity. Activities that cause degradation of soil structure include agricultural operations that may fragment aggregates and create compacted plow layers, compaction by heavy farm implements, and trampling by grazing animals (e.g., Hamza & Anderson, 2005; Nawaz et al., 2013; Stewart et al., 1990). Wetting-drying and freezing-thawing cycles induce swelling-shrinking effects in the soil and crack it. Biological activity plays a major role in soil structure formation and stabilization (Dexter, 1988; Oades, 1993). Earthworms and plant roots penetrate the soil and create biopores that provide preferential pathways for water and gas and help plants to proliferate their roots (Bottinelli et al., 2015; Bouchand et al., 2009; Colombi et al., 2017; Jarvis et al., 2016; Kroener et al., 2014). Root exudates, bacterial fuselage, and earthworm casting largely contribute to stabilization of soil structure (Oades, 1993). Soil structure generation and stabilization processes are slow and may take decades to centuries (Håkansson & Reeder, 1994; Webb, 2002). The large disparity between the characteristic time scales of degrading processes (rapid compaction at the scale of seconds) and the exceedingly long regenerative processes (years to decades) has contributed to misconceptions regarding the nature of the damage. This is exemplified by contrasting the intensive efforts in quantifying compaction with the limited attention given to mechanisms of recovery during which the main damage and loss of productivity occur (Keller et al., 2017). This bias propagates to reasonable characterization of compaction but virtually no measurements or metrics for soil structure recovery.

Quantifying soil structure noninvasively in space and time remains a challenge, involving the following four aspects: (i) what are the soil properties that best represent soil structure? (ii) how can we obtain information about these properties at the plot and field scale? (iii) what is the characteristic spatial scale and variability of these properties? and (iv) how do these properties evolve over time? Studies suggest that properties that capture the *soil structural form* (defined here as the pore size distribution, pore connectivity, and pore stability) are important in many dynamic soil processes (Keesstra et al., 2012; Naveed et al., 2016; Rabot et al., 2018; Stewart et al., 1990). Laboratory measurements and imaging capabilities allow for detailed quantification of soil structure within individual soil samples (Helliwell et al., 2013; Schlüter et al., 2014). Such descriptions of soil structure are, however, obtained under laboratory (not in situ) conditions. Furthermore, a major concern is that spatial and temporal undersampling is inevitable, which implies a limited capacity to infer spatial and temporal variations of soil structure and associated functions under natural conditions. In addition, there is limited knowledge on the potential bias occurring when findings from laboratory studies are extrapolated to in situ conditions. Apart from inherent limitations of extrapolating soil structural information from point measurements, certain aspects of the system dynamic responses to rainfall or other forcings become observable only at certain scales (profile, plot, and catchment). Visual soil evaluation methods (e.g., Guimarães et al., 2017) provide alternative means to examine the spatial variability of soil structure at the profile scale, yet they are subjective, empirical, highly invasive, and incapable of addressing soil structure dynamics. Considering the limitations of traditional characterization methods with reliance on point values and snapshots in time, we seek to expand the range of tools available for soil structure characterization at plot and field scales and across long time horizons by exploring the capabilities of geophysical methods to pick up signatures associated with structural features (rather than bulk properties).

In applied geophysics (e.g., Telford et al., 1990) and hydrogeophysics (e.g., Hubbard & Linde, 2011), measurements of geophysical fields are used to infer spatial variations in the physical composition of the Earth in order to delineate geological boundaries; identify deposits of minerals, oil, and gas; track the extent of groundwater and contaminants; and so forth. The analysis and interpretation of these upscaled geophysical-property models often relies on petrophysical relationships that link the inferred bulk geophysical properties with hydrological, transport, and mechanical properties of interest (e.g., Lesmes & Friedman, 2005; Mavko et al., 2009). Most such relationships were developed for consolidated porous rock formations, but their application has been widely extended to unconsolidated formations and soils. Efforts have been made in the last two decades to develop and standardize the use of geophysics for mapping soil properties (e.g., porosity, density, and clay content) and state variables (e.g., water content and salinity). For instance, applications of geophysical methods to agricultural planning and management (farm and field scale) are extensively discussed in compilations by Samouelian et al. (2005) and Allred et al. (2008). The European Soil Data Centre published a series of reports containing detailed methodologies to systematically map soil properties using geophysical data (Besson et al., 2010; Grandjean, Bitri, et al., 2009; Grandjean, Cousin, et al., 2009). Geophysical methods have been used to delineate soil horizons (Besson et al., 2004; Muñiz et al., 2016; Tabbagh et al., 2000), monitor water content (Garré et al., 2011, 2013; Michot et al., 2003), map soil texture (Grote et al., 2010; Sudduth et al., 2005), and characterize the effect of tillage on soil properties (Jonard et al., 2013). Most studies focus on the estimation of soil bulk properties, which mask the soil structural form features that are critical for soil functioning. The application of geophysical methods to characterization of soil and near-surface fluxes seldom consider dynamic changes in the pore spaces in measurement interpretation (i.e., the soil domain is considered constant and unaltered).

Several studies have attempted to characterize soil compaction through effects of soil bulk density on changes in Direct Current-resistivity (DC resistivity) (Besson et al., 2013), by studying reflections of electromagnetic (EM) waves from soil compacted layers using Ground-Penetrating Radar (GPR) (André et al., 2012; Muñiz et al., 2016; Petersen et al., 2005; Wang et al., 2016) or by relating inferred seismic velocities to soil strength obtained from penetrometers (Donohue et al., 2013; Keller, Carizzoni, et al., 2013). The motivation of most of the mentioned studies is grounded in the knowledge of how geophysical properties (e.g., electrical resistivity, dielectric permittivity, and seismic velocities) respond to variations of soil bulk attributes (e.g., clay content, density, and saturation). The impact of soil structural form properties (e.g., macroporosity and its connectivity) on geophysical data has not been considered systematically. Moreover, most laboratory experiments use soil samples that have been repacked (e.g., Z. Lu et al., 2004; Seladji et al., 2010), thereby removing or suppressing the natural soil structure. Only a few geophysical studies were devoted to investigate the sensitivity to soil structural form properties. For example, the detection of macropores using electrical measurements was studied by Moysey and Liu (2012), and the identification of preferential flow with electrical methods was reported by Koestel et al. (2008) and Garré et al. (2010).

Recent reviews (Binley et al., 2015; Jayawickreme et al., 2014; Parsekian et al., 2015) describe how geophysics can be used to gain information about subsurface functions and processes in the critical zone. However, the scope and discussions in these reviews are general to bulk behavior of the shallow Earth (~100 m). Our review seeks to explore how geophysical techniques could complement traditional soil structure characterization techniques, help to obtain insights about soil structure and its dynamics, and offer integrative ways of studying soil structure noninvasively at larger scales. For this, we seek to address three fundamental questions:

1. Are geophysical properties capable of providing information about structural features of soils beyond the already acknowledged links with soil bulk properties?
2. What is the expected sensitivity of measured geophysical data to relevant soil structural properties?
3. How to best combine different geophysical methods to obtain information on soil structure?

To address these questions, we set the following objectives:

1. To identify suitable geophysical methods that provide information about soil structural traits;
2. To review past research regarding the correlation between geophysical responses and soil properties and the applications that have resulted from those relations;
3. To devise observation strategies and ways to establish theoretical links between detectable changes in geophysical properties and soil structure; and
4. To propose an outlook of how geophysical observations can be used in combination with other measurements in order to quantify soil structural properties and dynamics.

2. Geophysics for Soil Structure Characterization: Concepts and Challenges

Descriptions of soil structure focus either on representation of secondary pore spaces not associated with texture (e.g., biopores) or, more generally, on the spatial arrangement, packing, and mechanical properties of the solid phase such as aggregates and compacted layers (e.g., Rabot et al., 2018). We will emphasize the pore space perspective due to its importance to soil ecological functioning (Rabot et al., 2018; Stewart et al., 1990), without neglect that compaction and aggregation are important soil structure components. The primary challenge is that bulk soil properties that are easy to measure (porosity or bulk density) offer limited insights about soil structure and functioning. To illustrate this, consider the schematic representation in Figure 1 (inspired by experimental observations presented by Keller et al., 2017), in which the soil texture in the three panels is identical and the bulk porosity is similar, the main difference is the different states of soil structure. Unfortunately, such nuanced views of structure related to soil pore spaces are difficult to quantify especially by geophysical methods that often cannot differentiate between soil traits of the different panels

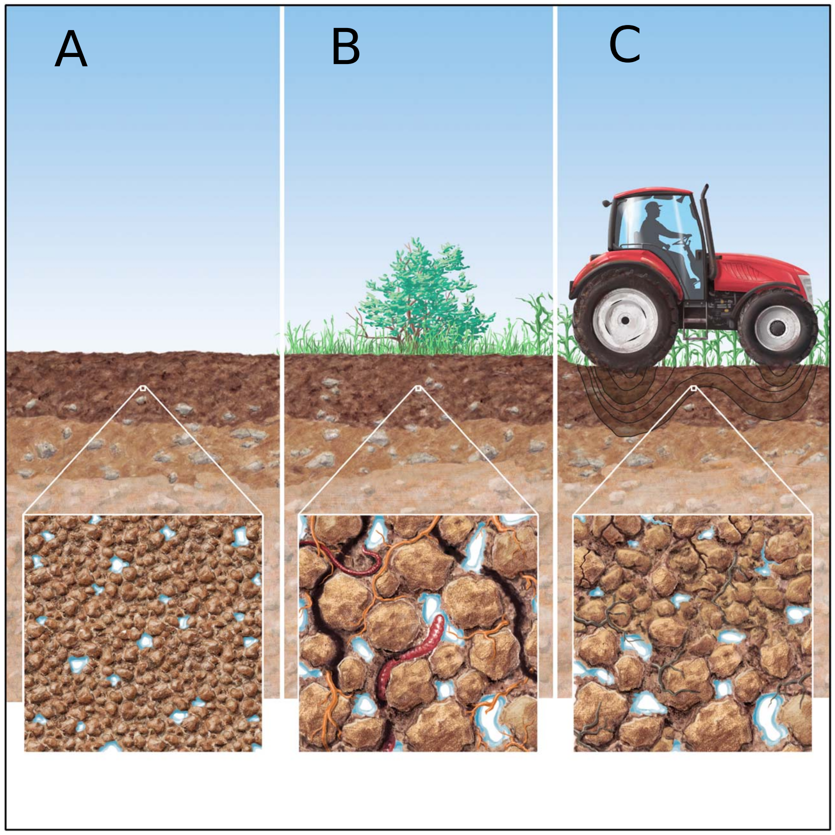

Figure 1. Schematic representation of soil structure along a transect. Each panel corresponds to a different soil structure: homogeneous soil (a); the same soil with secondary biologically induced structure (b); and soil structure from (b) as damaged by compaction (c).

in Figure 1. Soils are likely to be treated as a homogeneous domain with no signatures of biological activity (our reference soil in Figure 1 depicted in panel a).

Soil structure is generally difficult to quantify, and selecting geophysical methods and measurement strategies for this task is not obvious. One approach is to capitalize on indirect effects of soil structure and select methods and observations sensitive to changes in soil structure (e.g., enhanced drainage rates from soils with extensive biopores relative to soil with no structure or compacted). Such contrasting properties are schematically illustrated in comparison of Figure 1a, which shows a homogeneous structure, and Figure 1b, where earthworms and plants form channels and biopores and microorganisms stimulated by roots may excrete binding agents and promote aggregate formation. Soil biopores are an important element that differentiates soil structures. Despite contributing only to a small fraction of the entire soil porosity, soil biopores exert significant influence on water and gas transport and on near surface hydrology and the mechanical environment for growing roots. For example, the biopores become preferential flow pathways for air and water, which may increase the overall saturated hydraulic conductivity of the soil by several orders of magnitude, thereby, increasing water and oxygen availability for plant roots. These channels may additionally facilitate formation of biological hot spots and regions of low mechanical impedance for roots to grow in. In Figure 1c, we observe how a well-structured soil has been degraded and compacted by the passage of farm implements. The stresses applied by the passing tractor have resulted in the collapse and disruption of the largest pores in the soil. This reduction in macropore volume leads to a concurrent decrease in saturated hydraulic conductivity and an increase in mechanical impedance to root growth. A key aspect that set the pore structure in Figures 1b and 1c apart from Figure 1a is the presence of heterogeneities in the pore network characteristics (e.g., structures of different sizes and tortuosities) that are clear in Figure 1b and diminished but still present in Figure 1c.

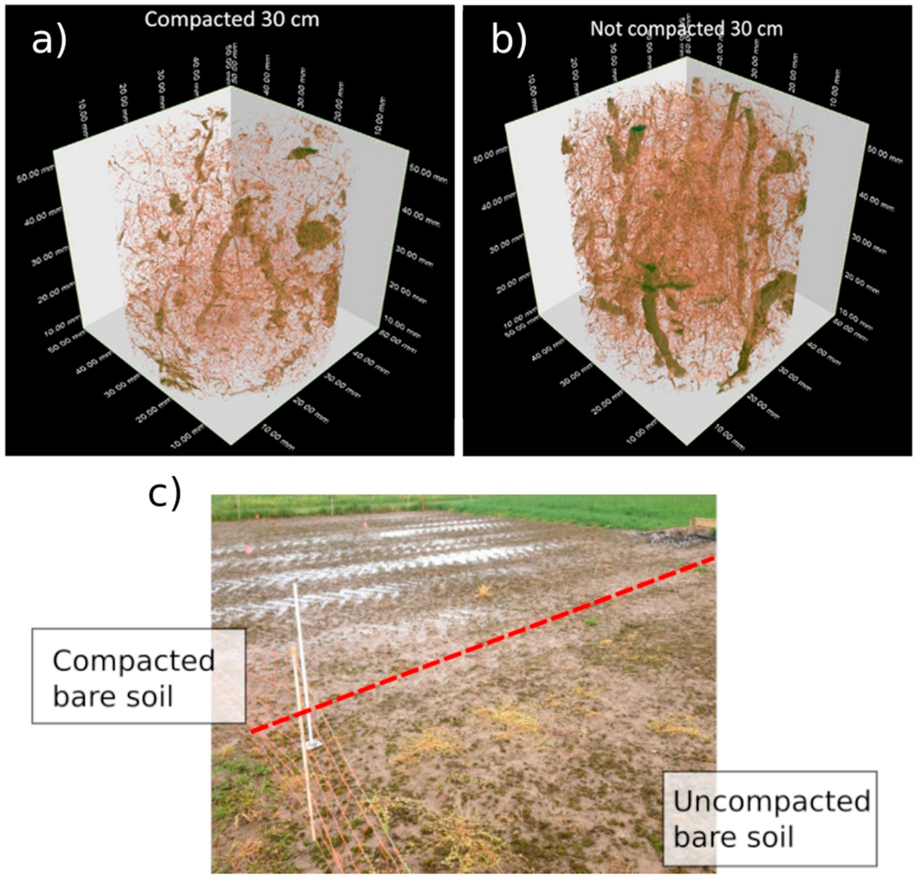

Figure 2. Soil pore structure detectable by microcomputed tomography (voxel size 60 μm, corresponding to a minimum pore width of 120 μm) of 100 cm^3 samples from (a) compacted and (b) uncompacted bare soil at 0.3-m depth sampled 2 weeks after the compaction event described by (Keller et al., 2017). The samples were taken from the field shown in (c). Figure modified from Keller et al. (2017).

The effects of soil compaction on macroporosity are illustrated in the X-ray computer tomography images of soil samples from compacted (Figure 2a) and uncompacted (Figure 2b) soil in the same field experiment near Zürich, Switzerland (Figure 2c; Keller et al., 2017). As the examples show, the reduction in overall porosities of the samples in Figures 2a and 2b is only 0.04 (from 0.44 to 0.4), whereas the hydraulic conductivity has been reduced by half (from 195 to 78 mm/hr), and the mechanical impedance measured by cone penetration nearly doubled (from 1.3 to 2.5 MPa). This example illustrates that compaction and subsequent soil structure recovery concern primarily soil macropores and the functionality they impart, and an important yet invisible ingredient, soil mechanical resistance. Our discussion of the impact of soil structure on geophysical signatures will be guided by this example where macroscopic properties routinely measured by geophysical methods show minute changes that do not capture the large impact on transport and mechanical behavior of the altered soil.

Considering the processes and interactions mentioned above, among the spectrum of geophysical methods, geoelectrical and EM methods with their inherent sensitivity to soil hydrological states are ideal candidates to assess the pore space and the influence of its different distributions on soil hydrology. Such methods include the DC-resistivity method (Binley & Kemna, 2005) targeting electrical conductivity, which depends strongly on pore space connectivity; the Induced Polarization (IP) method (Kemna et al., 2012) that senses capacitive properties that depend on the pore size distribution; and the GPR method (Annan, 2005) that is sensitive to soil moisture, interfaces, and cavities that could be associated with large roots or compacted layers.

Geoelectrical and EM methods, however, offer limited insights into the soil mechanical status. Soil mechanical properties (e.g., strength and elastic moduli) are better probed and characterized using shallow seismic methods (Donohue et al., 2013; Foti et al., 2011; Keller, Carizzoni, et al., 2013; Socco et al., 2010). The sensitivity of seismic measurements to the mechanical states of the soil and wave interactions with inclusions offer opportunities for detection of compacted layers, aggregation, and potentially large pores beyond what geoelectrical and EM methods provide. There is a wealth of experience and literature from geotechnical engineering on linking seismic signatures to soil mechanical states such as liquification resistance, penetrometer mechanical impedance, shear and bulk moduli, and soil density (e.g., Bhowmick, 2017; Mandal et al., 2016; Sabba & Uyanik, 2017; Yunmin et al., 2005). Additionally, cutting edge research into characterization of carbonate rock with vuggy pores (e.g., Skalinski & Kenter, 2014) offers a potential for using similar seismic measurements and methods for mapping large macropores.

A key step in the interpretation of geophysical methods is to define the links (and expressions) between the geophysical properties sensed and the soil properties and states of interest. In general, effective geophysical properties of a given heterogeneous volume composed by a multiphase porous medium (e.g., a soil) depend on two aspects: (i) a constitutive aspect, the geophysical property depends on the relative volumetric proportions of the constituents and their individual physical properties; and (ii) a structural aspect, the geophysical property depends on the way in which the different constituents are spatially distributed in the volume and how they connect. Virtually all theoretically based petrophysical models targeting electrical and mechanical properties were developed assuming an underlying structural model. However, emphasis in agricultural geophysics has been on estimating proportions without always recognizing the impact of the structural model that is embedded in a given petrophysical model. Application of geophysical methods for the study of soil properties and structure hinges on the use and development of petrophysical models that consider changes in pore spaces, are sensitive to soil constituents, and account for structural features.

Intuitively, it is expected that a high seismic velocity is an indicator of a stiff soil rather than a loose soil and that rapid rain infiltration as monitored by electrical methods suggests a well-connected pore network and the presence of macropores. Such qualitative expectations exemplify the already discussed potential for using various geophysical measurements in space and time to infer soil structure. A major challenge is, however, the lack of a systematic quantitative approach in which geophysical methods are used to capture such signatures and interpret soil structure. In this context, advancing soil structure characterization requires a set of ingredients (some existing and some that require further development) that can be summarized as (1) a set of geophysical methods that can sense the soil structural form (geoelectrical and EM methods) and the mechanical behavior (seismic methods), (2) a set of interpretation tools that focus on the signatures of the structural aspect of the geophysical response or property, and (3) a framework of survey configurations, combination of methods, and monitoring strategies that allow the inference of soil structure.

Soil structure information is important at the small scale of a plot or a field and at larger scales. For example, the parameterization of land-surface model often relies on the use of *pedotransfer functions* that relate soil attributes (often soil texture only) to hydraulic parameters. Recent studies have advocated for the urgent to include soil structure that could significantly affect infiltration and runoff in ways not predicted by soil texture (Hirmas et al., 2018; Or et al., 2013; Vereecken et al., 2016). In addition, the trend of agricultural intensification and associated adverse impacts on soil compaction and structure are expected to affect food security (Zhang et al., 2007). There is growing recognition for the importance of improving soil structure representation in Earth system models; geophysics may offer a critical role in providing such information at scales larger than the traditional point or sample scale measurements.

3. Geophysical Methods in Soil Science

In the previous section, we described the general challenges and possibilities with soil structure characterization using geophysical methods. Here we introduce selected studies that address soil properties using geophysical data. We discuss theoretical and empirical petrophysical relationships and supporting experimental evidence, as well as their use in a variety of soil science applications. The present literature review is not exhaustive, and we focus primarily on applications relevant to soil structure characterization. For more detailed introductions to the geophysical methods discussed herein, we refer to Binley and Kemna (2005; DC-resistivity and IP), Doolittle and Brevik (2014; EM induction [EMI]), Annan, (2005; GPR), and Steeples, (2005; seismic methods).

3.1. DC-Resistivity Method and EMI

The DC-resistivity method is a method that measures spatially distributed voltages resulting from current injections throughout an array of electrodes typically arranged on the soil surface or in boreholes. EMI

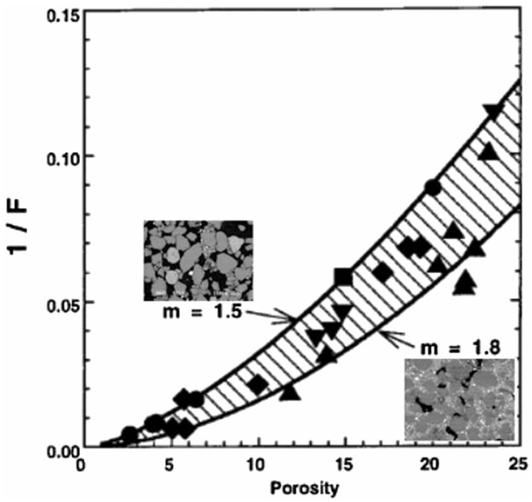

Figure 3. Relationship between the inverse of the electrical formation factor and porosity in Archie's law (equation (1)). Two values of the cementation exponent are shown to illustrate how it affects the formation factor: $m = 1.5$ has a more connected pore network than $m = 1.8$ as shown in the photographies, where the pore space is black and the solid phase is gray. Modified from Revil and Cathles (1999).

methods measure selected components of an EM field forming in the soil by induction in response to a prescribed EM field. Characteristics of the measured voltages or the induced EM field can be linked to subsurface electrical resistivity (Telford et al., 1990). DC-resistivity and EMI are addressed together because they both respond to electrical resistivity (ρ; or conductivity [σ], its inverse) although the underlying physical principles and sensitivity patterns are quite different.

Many authors (e.g., Corwin & Lesch, 2003; Friedman, 2005; Rhoades et al., 1976; Samouelian et al., 2005) discuss how soil electrical resistivity depends on the constitutive and structural aspects of soils that are captured by soil properties (e.g., bulk density and clay content) and state variables (e.g., soil salinity, water content, and water saturation), their interactions, and spatial arrangement. Significant research involving laboratory and field experiments has focused on the correlation between soil resistivity and one or more of these soil attributes. A suitable starting point is to combine the two experimental relations by Archie (1942) to express the impact of partial saturation on bulk electrical conductivity (σ) as

$$\sigma = \phi^m S_w^n \sigma_w = \frac{1}{F} S_w^n \sigma_w, \quad (1)$$

where $F = \phi^{-m}$ is known as the electrical formation factor and quantifies the increase in resistivity of the porous volume due to the presence of the solid matrix (assumed an insulator). The dependence on the constitutive properties is given by the electrical conductivity of the pore fluid (strongly linked to the salinity) σ_w, the saturation S_w, and the interconnected porosity (porosity sensed by electrical current flow) ϕ. The cementation exponent m and the saturation index n account for the contribution related to the soil structural form. The parameter m, for example, can in combination with ϕ be used to predict tortuosity (Nelson, 1994). The physical meaning of the cementation exponent m was discussed by Glover (2009), in which the inverse of the formation factor is interpreted as the connectedness of the pore network. A higher value of m indicates a reduction in effective pore connectivity as exemplified by Figure 3 and is related to the geometry of solid particles (Friedman, 2005). In most published field studies, m and n are treated as known constants or as fitting parameters.

A major shortcoming of equation (1) is that it ignores surface conductivity, which is significant in all soils and becomes an increasingly important contribution to bulk electrical conductivity with increasing clay content (Revil et al., 2017). Multiple petrophysical models account for surface conductivity in saturated media (e.g., the empirical Waxman-Smits model, Waxman & Smits, 1968), in which surface conductivity (σ_s) acts in parallel to the conduction paths in the pore space (the so-called high-salinity limit). Other models are based on Effective Medium Theory (EMT) (e.g., Bussian, 1983). EMT models are preferred because they do not make restrictive assumptions implying that the pore and surface conductivity electrical current pathways are parallel. Sen (1997) used the EMT framework to discuss the effect of different pore geometries and pore sizes on the electrical resistivity of rocks. In addition to the early work by Waxman and Smits (1968), models accounting for unsaturated conditions and surface conductivity (σ_s) have been proposed. For example, the model proposed by Linde et al. (2006):

$$\sigma = \frac{1}{F}\left[S_w^n \sigma_w + (F-1)\sigma_s\right], \quad (2)$$

was derived by volume averaging in the high-salinity limit. Other models can be found in Wunderlich et al. (2013) and Cosenza et al. (2009). Early models used to interpret electrical conductivity in soil science were often empirical (Rhoades et al., 1989). Regardless of the considered parameters (and rules) that account for different structural forms, none of these petrophysical models considers explicitly the role of different types of heterogeneities in the pore network (e.g., biopores) nor makes a differentiation between pore sizes (e.g., microporosity and macroporosity).

Field and laboratory evidence of correlations between electrical resistivity and soil attributes abound. Note that some of the clear links presented below for well-controlled laboratory experiments are not easy to demonstrate under field conditions when soil moisture, pore properties, and pore water conductivity change continuously and simultaneously. For example, McCarter (1984) measured the decrease in bulk resistivity with increasing water saturation by gradually compacting soil samples. The results suggest that the resistivity decreases with sample compaction. This is explained by increases in water saturation and the growing contribution from surface conductivity. However, the relative change in resistivity caused by soil compaction is expected to be soil dependent as many variables controlling electrical resistivity (e.g., porosity, pore connectivity, saturation, and volumetric clay content) vary with compaction. For example, for the well-structured agricultural soils shown in Figures 1b and 2b, compaction reduces the percentage of macropores and its connectivity but increases the contribution of surface conductivity. At a given partial saturation (implying unsaturated macropores) and salinity, the overall effect of compaction may, thus, be a decrease in the electrical resistivity of the soil. The work by Doussan and Ruy (2009) is intimately linked to soil structure. They demonstrate how unsaturated hydraulic conductivity can be predicted from electrical resistivity measurements at partial saturation by relating the pore diameter in the capillary equation with the characteristic pore diameter (proportional to the square root of the hydraulic conductivity) that results from applying percolation theory to porous media. Although their method requires exhaustive laboratory work (measurements of saturated hydraulic conductivity, electrical conductivity, and clay content), their estimations of hydraulic conductivity are in good agreement with independent measurements. Laboratory experiments by Moysey and Liu (2012) demonstrated that the apparent electrical resistivity of samples decreased by 30% when adding 4% of saturated macroporosity (generated by removing rods and saturating the resulting pores). They proposed theoretical bounds for bulk electrical resistivity by considering the macropores as cylindrical tubes that fully penetrate a homogeneous soil matrix. They demonstrate that the relative effect of macropore activation on bulk resistivity depends on the ratio between the resistivity of the pore fluid and the bulk resistivity of the background soil. Note that the condition of *fully saturated macropores* may seldom apply to a real soil.

Several studies have observed a signature of soil compaction (e.g., decrease in macroporosity, increase in density, and change in hydrological functions) on electrical resistivity. For example, Michot et al. (2003) calibrated relationships between water content and resistivity for three different soil horizons and used them to obtain water content sections from DC-resistivity tomograms. After irrigating the field in the first day of the experiment, they monitored the water content sections for 10 days and attributed low temporal variations of water content to zones that had undergone compaction. They discuss how this could be attributed to a reduction in hydraulic conductivity; this is in agreement with results presented by Richard et al. (2001). As mentioned above, laboratory experiments by McCarter (1984) showed that soil resistivity decreases in response to an increasing effective saturation due to compaction or increasing water content (especially for low moisture). These effects have also been captured in the field. Besson et al. (2004), for instance, studied the correlations of bulk density and electrical resistivity. They performed laboratory and field experiments to demonstrate that electrical resistivity is inversely correlated to changes in bulk density and, based on their observations, concluded that the DC-resistivity method has potential for delineating structural features such as the plow pan, wheel tracks, and compacted soil lumps. The evolution of soil resistivity following compaction was studied in a 1-year-long field experiment by Besson et al. (2013). Along with DC-resistivity, they monitored soil water content and temperature and performed periodic measurements of density via soil coring in compacted and noncompacted soils. Their results show that effects of compaction in electrical resistivity, attributed to changes in bulk density and water content, persisted after 1 year of monitoring. Rossi et al. (2013) studied the effect that tillage has on DC-resistivity data. Their setup contained DC-resistivity profiles located perpendicular to a plot with an alternating tillage pattern (Figure 4a); high resistivity anomalies are observed in the tilled zones. They suggested that the electrical heterogeneity was brought by soil breakup due to tillage, yet they emphasized the difficulty in recognizing effects of tillage as they were to some extent masked in the natural electrical variability of the soil. Clearly, this type of effects is best studied using a reference baseline prior to a perturbation (e.g., tillage or compaction). Figure 4b shows changes in electrical resistivity of a soil following compaction (see also Keller et al., 2017) with electrical resistivity dropping up to 15% in the compacted zones, which in this study was primarily attributed to increases in soil surface conductivity. It must be highlighted, however, that uncertainty in time-lapse DC-resistivity measurements increases when soil structure changes occur due to agricultural machinery, since the process of removing and replacing the electrodes will introduce position changes that may impact the data and subsequent inversion results. As discussed previously,

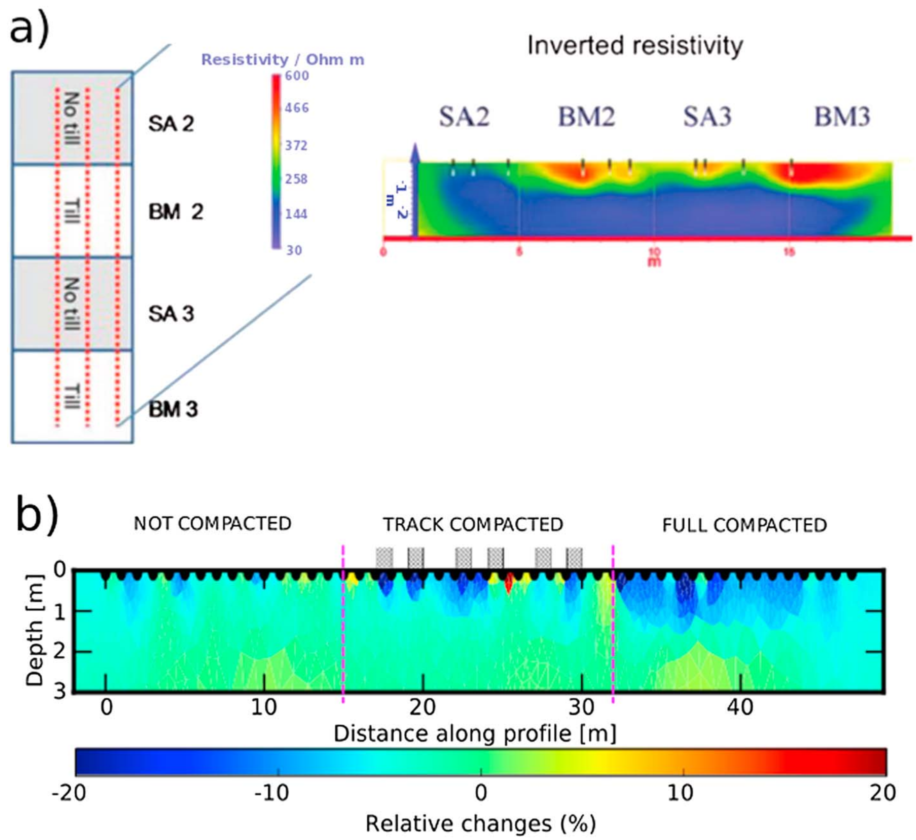

Figure 4. (a) Experimental setup of Direct Current-resistivity measurements after the tillage experiment by Rossi et al. (2013), soil electrical resistivity tomogram from data taken on one of transects in that cross the alternating tillage (BM2 and BM3), and nontillage (SA2 and SA3) management regions (reprinted from Rossi et al., 2013). (b) Relative changes (before and after soil compaction) in inferred electrical resistivity of a soil (Keller et al., 2017). Gray rectangles indicate track passages, and the full compacted part was made by consecutive passages of tracks next to each other.

changes in soil structure introduced by compaction and tillage are expected to affect electrical resistivity, and current petrophysical models (e.g., equations (1) and (2)) may offer limited means to interpret these signatures.

Daily and seasonal variations in soil temperature are important to consider due to their potential effects on electrical resistivity (+1 °C may yield 2% decrease in electrical resistivity) and dielectric measurements by GPR and Time Domain Reflectometry (TDR). To account for thermal effects on geophysical measurements, a correction is applied to measured electrical resistivity or dielectric permittivity (ϵ) values, these corrections are discussed in Besson et al. (2008) and in Or and Wraith (1999).

Considering the current state of the literature, we consider soil electrical resistivity as a valuable property for studying soil structure. It is one of the most commonly used geophysical properties in soil investigations, and it carries information about a wide range of soil properties. This multitude of sensitivities can be seen both as an advantage and a disadvantage. We identify that there is a need to extend existing petrophysical relationships to consider aspects related to soil structure (e.g., macroporosity) together with associated laboratory experiments. Existing methodologies for estimating electrical resistivity (survey configuration, inversion techniques, etc.) are well developed and adapted to different spatial scales.

3.2. IP

The IP method is an extension of the DC-resistivity method. This method has a strong potential for soil structure characterization and deserves further attention, given that the mechanisms (e.g., polarization of grains and pore throats) that govern the IP responses occur mainly at the interface between the pore space and the solid matrix (Kemna et al., 2012). This should imply that IP properties are strongly linked to the structural aspect of the studied soils. For example, in the context of groundwater studies, Slater (2007) provides a discussion about the value of IP properties to estimate hydraulic conductivity in aquifers. In time-domain IP, the main geophysical property targeted is the chargeability ($M = V_s/V_p$), defined as the ratio between the secondary voltage (voltage immediately after the current is shut off, V_s) and the primary voltage V_p (Binley & Kemna, 2005). In spectral IP (SIP), the frequency dependence of the complex electrical conductivity ($\sigma^* = 1/\rho^*$) is measured by applying alternating currents (Kemna et al., 2012). IP properties are very sensitive to the specific surface area of the soil, which is determined mainly by the clay content and clay type. Indeed, IP measurements complement DC-resistivity measurements by constraining the contribution of surface conductivity to electrical conductivity. As mentioned above, the IP method is an extension of the DC-resistivity method, and measurements can be made using a typical four electrode configuration. Yet obtaining high quality IP data is technically considerably more difficult since cable shielding and special electrodes need to be used to avoid capacitive coupling overruling the polarization effect from the soil.

Rather than studying the chargeability, authors have often focused on chargeability normalized by the electrical resistivity: the so-called normalized chargeability ($MN = M/\rho$), which is strongly correlated with the clay content of the soil (e.g., Börner et al., 1996). Interestingly, the chargeability and resistivity do not carry conclusive information on soil clay when considered separately. The capability of the normalized chargeability to discriminate zones was studied by Slater and Lesmes (2002). They observed a clear lithological anomaly in their MN sections that was not present in their ρ and M sections. Thus, it is preferential to work with the normalized chargeability as it generally presents a linear dependence on the cation exchange capacity and clay per unit of volume (Weller et al., 2013). The time-domain IP method is comparatively underused in soil science, partly because of the perceived difficulty to obtain high-quality field data (low signal-to-noise ratios and EM coupling effects) and the relative lack of predictive and robust petrophysical models that link IP parameters to soil properties. Recent laboratory studies on soil samples by Revil et al. (2017) are expected to promote its use.

We argue that SIP parameters have a great potential to offer insights about the structural form of the soil. The Cole-Cole model, for example, is an empirical model that links the frequency dependent complex electrical conductivity to a relaxation time (often assumed to depend on either the grain size or the pore size) and an exponent that depends on the width of the grain size distribution (Friedman, 2005). Qualitatively, Ghorbani et al. (2008) observed the effect of macroporosity on the complex electrical conductivity. In an infiltration experiment, they observed a drop in the phase of the complex conductivity that coincided with the time at which the macropores became saturated by the water front (as measured by a tensiometer). This *two-stage* behavior of the phase data was interpreted as a decrease in the net polarization produced by the diffusion of ions from wet aggregates to the saturated macropores as they were filled with water. Based on this, they suggested that SIP parameters could be used to monitor the saturation and desaturation of macropores in soils. This result is in good agreement with laboratory work by Breede (2013). Here the measured variations with saturation on SIP properties in repacked soil samples (mixtures of sand and clay) were qualitatively attributed to polarization effects related to saturation and desaturation of heterogeneous pore spaces.

The SIP method is still under development; the current state-of-the-art is reviewed by Kemna et al. (2012). The experimental evidence addressed in this section suggests that future developments and improved understanding of polarization properties of structured soils will eventually give IP methods a central role for characterization of the structural form of soils.

3.3. GPR

GPR is a high-frequency EM method that is governed by a wave equation. In the typical setup, an EM pulse is transmitted by a transmitter antenna placed in contact with the surface of the soil, and the response is measured over time using a second receiving antenna placed at a fixed distance. The system is moved along a profile while repeating the measurements with a short measurement interval (for more details, please refer to Annan, 2005). The GPR response is given by the interaction of EM waves at the soil-air interface, its propagation through the soil and scattering at interfaces. At GPR frequencies, the propagation velocity (v) and the reflexion coefficients of EM waves depend mainly on the relative permittivity (so-called dielectric constant, $\kappa = \epsilon_{\text{medium}}/\epsilon_{\text{vacuum}}$) of the medium in which it is traveling following the relation:

$$v = \frac{c}{\sqrt{\kappa}}, \tag{3}$$

where c is the speed of light in a vacuum. The dielectric constant of water ($\kappa \approx 80$) is widely different than the dielectric constant of both the matrix minerals ($\kappa \sim 5$) and the air ($\kappa \approx 1$). For this reason, the GPR propagation velocity is strongly dependent on the water content, and the scattering of GPR signals is mainly caused by variations in water content.

Some of the most widely used models that relate the dielectric constant to the soil constituents are purely empirical. For example, the Topp model (Topp et al., 1980), in which the predicted dielectric constant depends only on the soil water content through a polynomial relationship, is given by

$$\kappa = 3.03 + 9.3\theta + 146\theta^2 - 76.7\theta^3. \tag{4}$$

More theoretically based parametrizations are preferred as they consider different soil properties and assume a certain structure. For example, the volume averaging relation proposed by Linde et al. (2006) connects the dielectric constant to soil properties by

$$\kappa = \frac{1}{F}\left[S_w^n\kappa_{\mathrm{w}} + \left(1 - S_w^n\right)\kappa_{\mathrm{a}} + (F-1)\kappa_{\mathrm{s}}\right], \tag{5}$$

where κ_{w}, κ_{a}, and κ_{s} are the dielectric constant of water, air, and the soil matrix, respectively, or the Lichteneker-Rother model (e.g., Roth et al., 1990):

$$\kappa = \left[\theta\kappa_{\mathrm{w}}^{\alpha} + (1-\phi)\kappa_{\mathrm{s}}^{\alpha} + (\phi-\theta)\kappa_{\mathrm{a}}^{\alpha}\right]^{\frac{1}{\alpha}}, \tag{6}$$

with typically $\alpha = 0.5$, more generally, $\alpha = 1/m$, where m is Archie's cementation exponent (Brovelli & Cassiani, 2008).

The applicability of GPR in soil-related field-based studies has been widely reported in the literature. For instance, Grote et al. (2003) derived a site-specific petrophysical relationship from soil samples and used it to investigate the accuracy of soil moisture estimations obtained from GPR surveys. The comparison between moisture estimations using GPR and sample measurements shows correlation coefficients as high as 0.98 for 900-MHz data and 0.92 for 450-MHz data. The travel time to a reflected layer measured by GPR was used by Lunt et al. (2005) to estimate water content in a Californian vineyard where they had certain knowledge about the presence of a reflective layer in the soil. They calibrated a site-specific relationship between dielectric constant and water content and used it to map water content. Krueger et al. (2013) used a combination of EMI and GPR to map soil depth and used these estimations as input in a grain yield model. The accuracy of the model used to predict grain yield was improved by using such a geophysically assisted approach. André et al. (2012) attributed GPR reflections in a transect across a vineyard to soil compaction. Zones with strong reflections present a compacted soil profile and a poor development of the vine in comparison to a weak reflection in an uncompacted zone where the vine presents a higher development. Di Matteo et al. (2013) discussed the effects of near-surface dielectric constant (with a sensitivity to depths on the same order of magnitude as the EM wavelength in the soil) on early time GPR amplitudes; these concepts were confirmed by numerical simulations. In fact, the dielectric constant is highly correlated to the amplitude of early GPR signals. These findings were used by Algeo et al. (2016) to map soil water content using GPR amplitudes in a field-scale experiment.

With respect to soil structure and compaction, Petersen et al. (2005) explored the value of GPR, EMI, and refraction seismics to assess soil structural changes caused by soil compaction. For the compacted soil, they observed strong reflections in GPR signals under humid conditions. The contrast in dielectric constant that was causing these reflections was attributed to layers of variable water content that were considered indicators of soil compaction. Wang et al. (2016) used GPR data to study the correlation between soil properties and GPR wave speeds in compacted soils (both laboratory and field). Their results show that the wave speed (i.e., also dielectric constant) is influenced by water content, bulk density, and penetration resistance; they did not provide an equation describing these relationships because of insufficient data. Figure 5 shows GPR data collected with an 800-MHz antenna at the experimental field (soil profile similar to Figure 5c) described by Keller et al. (2017). The data were collected in a soil transect at two stages: before compaction (Figure 5a) and after (Figure 5b) the compaction of certain zones by a passing tractor. Note that the water content at the times of measurements of Figures 5a and 5b was similar in the noncompacted treatment. However, the water content of soil transects that underwent compaction is expected to be lower. A zero time correction was applied to the data, followed by DC-shift filter and a linear gain function. It is possible to identify the compacted zones based on features that appear in the data. For example, the postcompaction radargram shows an enhancement of signal amplitudes over the soil regions that underwent compaction.

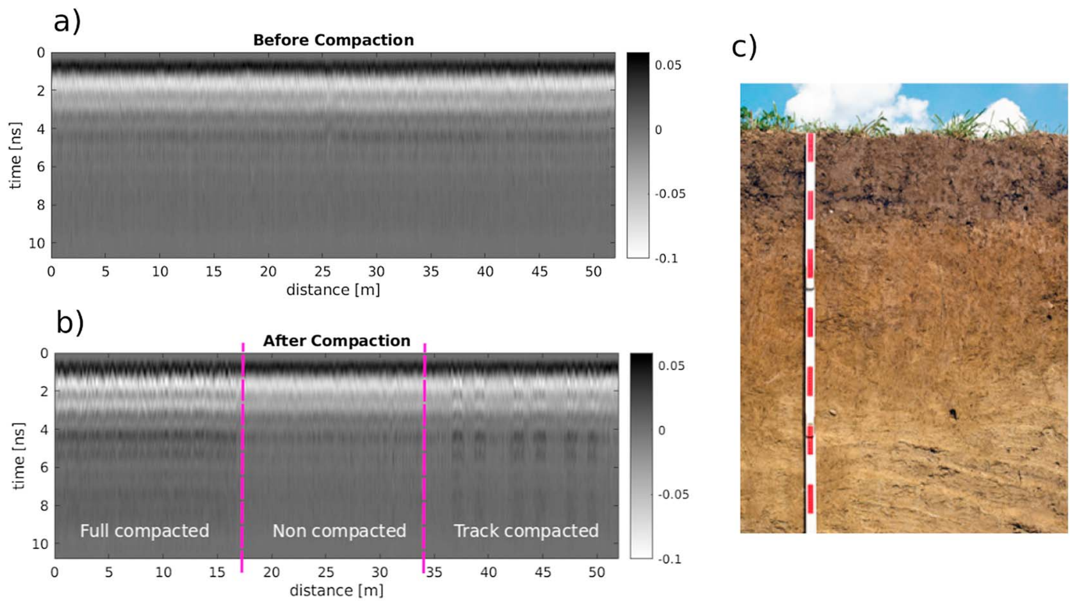

Figure 5. Ground-Penetrating Radar spatial identification of compacted zones in a compaction field experiment (described by Keller et al., 2017) in Zürich, Switerland. (a) Data collected before the compaction event. (b) Data collected after the compaction event at the same location. (c) Soil profile from a noncompacted zone in the experimental field. The red and white colors in the ruler alternate every 10 cm.

The main value of the GPR method is its ability to estimate water content at a high spatial resolution (Klotzsche et al., 2018). It offers estimates of soil moisture that are useful to account for dynamics of water losses (evaporation and root water uptake). Furthermore, it has have the potential of providing information related to natural layering and layering introduced by compaction. Survey designs and acquisition systems are relatively well developed, but readily available and accurate techniques for quantitative interpretation of GPR data are more limited. A major limitation of the GPR method in soil science investigations is that the signal is highly attenuated in electrically conductive soils (high clay content).

3.4. Seismic Methods

Seismic methods involve the measurement of ground displacement velocity (or acceleration) generated by compressional and shear waves produced by (most often) an artificial source (e.g., dynamite explosion or hammer impact). The use of seismic methods in soil science studies is not as common as electrical or EM methods. Nevertheless, seismic wave fields contain information about the mechanical properties of the subsurface and may offer insights about soil structure that other geophysical methods cannot provide (Keller, Lamandé, et al., 2013).

For a homogeneous media, seismic velocities (pressure wave v_p and shear wave v_s) are related to elastic moduli and density through

$$v_p = \sqrt{\frac{K + 4/3\mu}{d}}, \tag{7}$$

and

$$v_s = \sqrt{\frac{\mu}{d}}, \tag{8}$$

where K is the bulk modulus (or compressibility modulus), μ is the shear modulus, and d is the density. The elastic moduli of a porous medium (especially soils) are complex functions of the individual properties of the components of the mixture. Various relationships exist that link elastic moduli to, for example, porosity (see Schmitt, 2015, for a review of the most common relationships). One of the most widely used are Hashin and Shtrikman (1963) relationships, which bound the elastic moduli of a mixture of grains and pores.

The bulk modulus of water is generally greater than that of the background dry matrix in unconsolidated soft materials like soils, so the impact of water content on the effective elastic properties of a soil is central yet not fully understood and a cause of controversy in the scientific community (e.g., Shin et al., 2016). The theory of elastic wave propagation through a multiphase porous material proposed by Brutsaert and Luthin (1964) is often used to study the influence of water content on *P* wave velocity. It predicts

$$v_p = \sqrt{\frac{0.306\, a\, p_e^{1/3} Z}{d\, \phi\, b^{2/3}}} = \sqrt{\frac{\frac{0.306\, a\, p_e^{1/3} Z}{\phi\, b^{2/3}}}{d}}, \tag{9}$$

where *a* and *b* are fitting parameters that largely depend on the elastic properties of the background dry matrix (Shin et al., 2016), p_e is the effective pressure that depends on the capillary pressure (indirectly related to water content by the soil characteristic curve), and Z is a function of the degree of saturation or, more specifically, of the effective bulk modulus of the mixture of fluids (water and air) in the pore space. Equation (9) describes the dependency of the *P* wave velocity on the saturation (encoded in Z) and effective pressure. In such a description, the conceptualization underlying the calculation of the effective bulk modulus of the fluid mixture plays a predominant role and may determine the trend of the relationships between *P* wave velocity and saturation. For instance, the effective bulk modulus will be very different if using arithmetic or harmonic (for which Z reduces to 1 for almost the entire range of saturations) weighted averages of the bulk modulus of the two fluids (Domenico, 1977). When measuring the *P* wave velocity over a certain spatial and temporal scale, such simple assumptions do not well explain the actual water phase distribution as it is affected by many processes (infiltration, drainage, and evapotranspiration), and it is very unlikely that the same mixing rules apply at all times (e.g., the empirical relation by Brie et al., 1995). There are also other effects of water content (or saturation) on the *P* wave velocity: indirectly by its impact on water potential and directly by its impact on density.

Z. Lu and Sabatier (2009) performed a 2-year-long field experiment to study the above-mentioned effects of saturation and matric potential on the *P* wave velocity by using Brutsaert and Luthin's theory. They instrumented a trench in a soil to measure *P* wave velocity, water content, and matric potential and back filled it with a soil mixture. They assume that $Z \approx 1$, which corresponds to a homogeneous distribution of the water in the pore space. Their results show that *P* wave velocity relates to matric potential in good agreement with the predictions by Brutsaert and Luthin (1964). They observed a decreasing of *P* wave velocity with increasing water content. Using equation (9), this counterintuitive result can be understood as a consequence of the reduction of the capillary force acting on the soil particles and (to a lesser extent) an increase of the bulk density, leading to a decrease in the nominator (bulk moduli) and increase in the denominator of equation (9), respectively. A more detailed discussion can be found in Shin et al. (2016). Brutsaert and Luthin's theory does not consider the relative movement between the solid and fluid phases, so dissipation of energy associated with such phenomena is ignored. Furthermore, the soil study by Z. Lu and Sabatier (2009) was performed on an artificial soil mixture, without a well-developed structural form; a situation for which it is reasonable to assume $Z = 1$ (c.f. Figure 1a). Soil aggregation in a well-structured soil and consolidation of the solid matrix should reduce the relative effect of the capillary forces on the bonding of particles. Flammer et al. (2001) measured *P* wave velocities and water content in undisturbed soil samples and observed a high sensitivity of the velocity to the distribution of water in the samples that was in turn influenced by preferential flow in macropores. For this reason, we expect Z to be a time-varying function, and the appropriate way to derive it for a heterogeneous water-air mixture is an open question.

Unlike in geotechnical applications and reservoir characterization, seismic methods are relatively underused in soil characterization for ecological and agricultural applications. We have mentioned the potential adaptation of some of the methods to characterize soil compaction and even soil structure by drawing analogies with petrophysical approaches used in characterization of carbonate reservoirs (Skalinski & Kenter, 2014). The Multichannel Analyses of Surface Waves has been widely applied in civil engineering studies (e.g., Park et al., 2002, 2007; Xia et al., 2000) to map shear wave velocities, known to be strongly linked to different soil mechanical properties (e.g., Foti et al., 2011; Socco et al., 2010). For example, the shear wave velocity has been related to the liquefaction resistance in soil samples by Yunmin et al. (2005). Sabba and Uyanik (2017) presented an empirical exponential relationship between the uniaxial compressive strength of samples of reinforced concrete and their shear-wave velocities ($R^2 \sim 0.90$). The principles of MASW methods have been successfully used for in situ studies, for example, the methodology employed by Ryden and Park (2006) to determine the thickness and shear velocities of pavement layers. These concepts have already been used in other systems

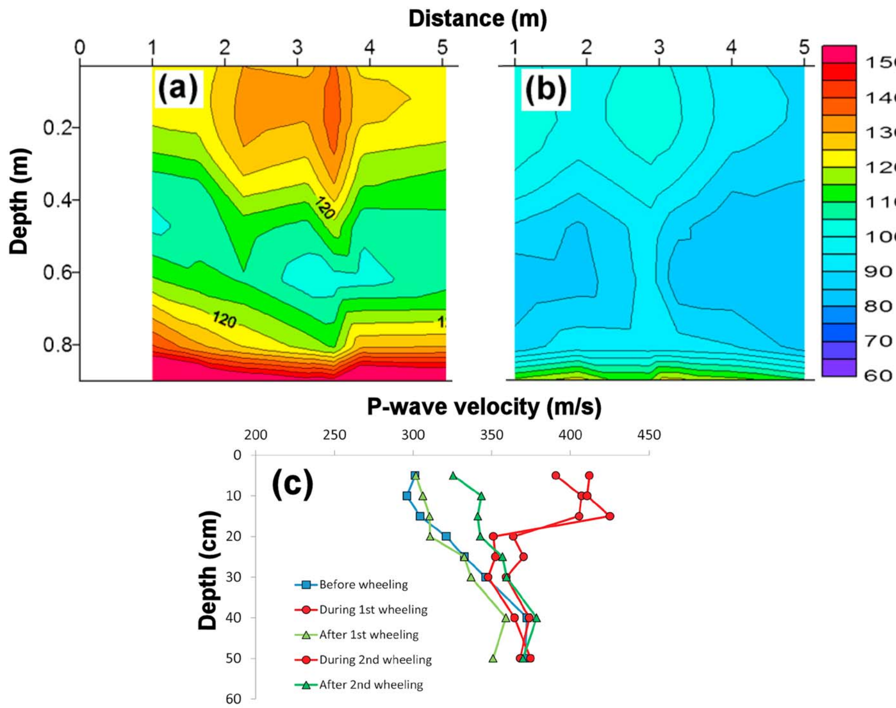

Figure 6. The sensitivity of seismic measurements to soil compaction and load: *S* wave sections from Multichannel Analysis of Seismic Waves deduced from a compacted zone (a) and a noncompacted zone (b), note lower *S* wave velocities for the noncompacted soil (from Donohue et al., 2013); and (c) *P* wave velocity with depth measured in a wheeling experiment marking changes during load application and after passage of the wheel (from Keller, Carizzoni, et al., 2013).

with similar characteristics (e.g., permafrost study by Dou & Ajo-Franklin, 2014). An interesting study for soil structure was presented by Mandal et al. (2016) for linking soil strength with *P* wave velocities. They measured flexural strength and *P* wave velocities of different types of soils that were stabilized with different binders. Their study shows that minute changes in soil bulk density may induce significant impact on seismic velocities. Similar effects are induced by soil biological activity with cementation and increase in strength and creation of stable aggregates.

Penetrometer mechanical impedance is an important soil trait that affect root growth and other aspects of soil biological activity (Ruiz et al., 2015). In a manner similar to artificial cementing experiments of Mandal et al. (2016), near-surface seismic measurements offer a relatively direct window to quantify soil stabilization by biological agents. Some authors have proposed to use estimations of seismic velocities in combination with parameters derived from penetrometer resistance measurements to estimate bulk density by using empirical relationships (Burns & Mayne, 1996; Mayne et al., 1999). For example, the dynamic resistance is related to the maximum shear modulus of the soil by Lunne et al. (1997) and Mayne and Rix (1993). Then, the density could be derived by the combination of these empirical relations with equation (8). More interestingly, seismic velocities have shown to be correlated with penetrometer resistance measurements, thereby, confirming the link between the seismic velocities and the mechanical impedance of soils. For example, Donohue et al. (2013) explored the possibility of using MASW for detecting soil compaction. They obtained a rather strong ($R^2 = 0.66$) correlation between inverted shear wave velocities and bulk density from sampled cores and penetration resistance taken in the field. The spatial distribution of the shear wave velocity is shown in Figure 6. Indeed, the soil in Figure 6a that was presumably compacted presents higher velocities and suggests the presence of a compacted layer in comparison with the soil in Figure 6b where a more homogeneous distribution of seismic velocities is observed. *P* wave velocities were inferred during a wheeling experiment (passing of agricultural machinery) by Keller, Carizzoni, et al. (2013). They inferred *P* wave velocities at different depths along a soil profile crossing the traffic line at various stages (before, during, and after wheeling). As shown in Figure 6c, there is a strong increase in velocity during the first wheeling, a relaxation to its initial value between the first and second wheeling, a similar increment in velocity during the second wheeling and a relaxation to an intermediate value after the second wheeling. This illustrates the strong influence of soil structure on seismic velocities, yet quantitative interpretation is challenging with the existing tools. They also measured penetrometer resistance and took samples at various depths to measure bulk density. Their results show a correlation between *P* wave velocity and bulk density, and they reported a site-specific relationship between *P* wave velocity and penetrometer resistance.

Despite the lack of a definitive understanding of the mechanisms that influence seismic velocities in soils, we argue that seismic methods are underused in studies of soil mechanical status and soil structure. In fact, the direct dependence of seismic waves on the soil mechanical properties makes seismic methods essential for soil structure evaluation, given that they can resolve mechanical states that are not observable by other (e.g., geoelectrical) geophysical methods.

3.5. Other Geophysical Methods

We have discussed the geophysical methods that we consider best suited to address soil structure characterization. One should keep in mind, however, that a wide repertory of geophysical methods exist, and the usage of some of them may be relevant in this context (Binley et al., 2015; Grandjean, Cousin, et al., 2009). For instance, the self-potential method (Revil et al., 2012) is a passive electrical method, in which naturally occurring electrical potential differences are measured. The streaming potential contribution that is related to fluxes and water saturation is important in soils (Doussan et al., 2002; Jougnot et al., 2012, 2015) and can be associated to root water uptake or percolation following precipitation. Laboratory-, borehole-, and surface-based nuclear magnetic resonance methods are sensitive to water content and, in some cases, to hydraulic conductivity (Behroozmand et al., 2014; Binley et al., 2015). Gravimetric measurements (especially time-lapse) can be useful to constrain estimates of hydraulic states and fluxes (Blainey et al., 2007). The hyperspectral method uses the light reflected by the soil, which in turn is determined by its composition (texture, organic matter, and salinity) and could be useful to improve estimates of related properties (e.g., clay content) over large areas (Ciampalini et al., 2015).

4. Opportunities for Geophysical Soil Structure Characterization

Following our review of geophysical methods and identification of current limitations for soil trait inferences, we evaluate next theoretical developments and applications of geophysical methods aimed specifically at resolving soil structural features directly or indirectly by examining the effects of soil structure on the soil system dynamics.

4.1. Petrophysical Models of a Structured Soil: Pedophysical Models

Petrophysical models are central to the interpretation of geophysical measurements. We have discussed some of their shortcomings for soil structure identification due to their reliance on volumetric proportions or neglect of key physical processes in their construction. A detailed treatment of the wide field that links geophysical responses with soil and geological material properties is beyond the scope of this review. We have presented several relationships for linking electrical conductivity to porosity (equations (1) and (2)), dielectric constant with water content (equation (6)), and seismic wave velocity with soil bulk density (equation (9)). However, the development of specific pedophysical relationships (associated to soils and not to rocks) is needed to include effects of soil structure on geophysical properties. A promising approach for incorporating soil structure is offered by EMT and related formalism (Berryman, 1995; Friedman, 2005).

In the absence of surface conductivity (a condition that hardly occurs in any soil), Day-Lewis et al. (2017) compared predictions of electrical petrophysical models that consider two porosity domains (Figure 7) with different salinities. One prediction is based on a simplified (assuming parallel conduction) weighted arithmetic average of the fluid conductivities, and the other is derived using a variant of EMT called differential effective media (DEM) theory (Bussian, 1983). Predictions made by the latter model exhibit a very good match to pore network simulations of dual-domain solute transport. In this formulation, each porosity domain has its own cementation exponent (hence connectivity), and the interaction between the two conducting domains are accounted for. Furthermore, there is a functional relationship imposed between the two connectivities, thereby, limiting the number of free parameters. A similar parameterization could be pursued for aggregated soils by differentiating between interaggregate and intra-aggregate pore space. Another example for which the shape and distribution of constituents has been studied extensively is dielectric mixtures (Friedman, 2005; Wunderlich et al., 2013) where the arrangement of the phases or the shapes of the elements in the mixture affect the effective value of the upscaled petrophysical property (Sihvola, 1999).

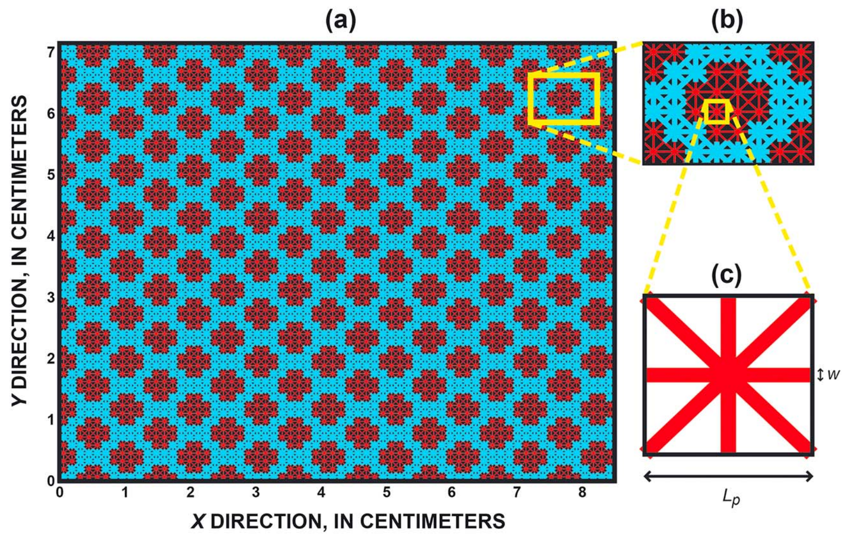

Figure 7. (a) Diagram illustrating a bimodal pore network model consisting of a pipe lattice with a given porosity (blue) between accumulations of zeolite grains (red) with (c) an internal porosity. From Day-Lewis et al. (2017).

The richness of EMT has been demonstrated by providing a rigorous basis for relationships that were proposed initially on empirical or ad hoc basis. For example, the Hanai-Bruggeman mixing formula derived from DEM (e.g., Bussian, 1983) reduces to Archie's law when considering that conduction is dominated by the solute (see also Sen et al., 1981). The exponent $m = 3/2$ results from considering spherical inclusions in the DEM formalism, and the different values of this exponent emanate from considering ellipsoidals with different eccentricities and orientations (Cosenza et al., 2009). As for the high frequency permittivity, Zakri et al. (1998) demonstrated that the Lichteneker-Rother formula (equation (6)) can be derived from EMT by considering a symmetric Bruggeman mixing rule in which ellipsoidal inclusions parallel to the applied field are used and whose eccentricities follow a beta distribution. The inclusion of EMT formalism for soil structure characterization would require additional parameters and render the inference in field settings more complicated and less unique. To address some of this complexity, we will discuss shortly (section 4.2) approaches that would harness concurrent measurements by different geophysical methods to better constrain the interpretation.

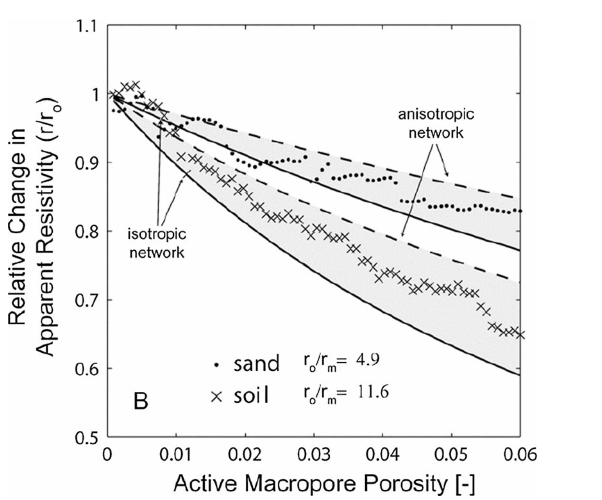

Figure 8. Relative change in apparent resistivity as a function of saturated macroporosity. Symbols show experimental data for true (x) and sand (·) soils. Shaded areas indicate the region between the theoretical bounds derived for isotropic and anisotropic macropore networks. Here r/r_0 is the ratio of the initial soil resistivity to the macropore resistivity (from Moysey & Liu, 2012).

Laboratory investigations by Moysey and Liu (2012) show sensitivity of the electrical conductivity to the saturation of macropores in soil samples. This work (shown in Figure 8) demonstrated that even slight increases in active (saturated) macroporosity can produce considerable decreases in electrical resistivity, and it proposed theoretical bounds by arithmetic and harmonic averages. These results illustrate that there is a measurable impact of macropores on electrical measurements and support our discussion that encourages the development of appropriate pedophysical models to capture such effects.

Mechanistic models describing IP responses are based on the assumption that the polarization process can be decoupled such that the contribution from each grain (or pore) can be summed up by a convolution operator. Furthermore, grain (and pore) polarization models are generally based on idealized geometrical shapes that might poorly describe actual pore structure. This might explain why the packing of identical matrix material has a strong IP effect, which cannot be captured by existing models (Bairlein et al., 2014; Kemna et al., 2012). To better understand the soil structure signature of the IP method, we suggest that laboratory work on undisturbed soil samples together with advanced soil structure imaging and modeling is needed.

Improving our understanding of the links between seismic properties and soil structure is an important and promising area that could enable noninvasive probing of the mechanical state of the soil by accounting for effects of structure on the seismic wave propagation. Changes in soil structure contain several invisible effects on soil function, primarily with loss of large pores and alteration of their topology (continuity), but even more subtle that these changes are the alteration of soil mechanical resistance and strength that affect biological life in soil. The impacts of compaction and load on seismic velocities have been demonstrated in several studies including the aformentioned wheeling experiment (Keller, Carizzoni, et al., 2013) where *P* wave velocities were sensitive to transient (load) and permanent changes in soil mechanical properties after wheel passage. The representation of such changes on seismic wave propagation hinges on advancing petrophysical modeling of commensurate changes in key variables (bulk density, soil stiffness, elastic moduli, and more). Markov et al. (2005) used DEM theory to study a related problem, namely, the sensitivity of effective elastic properties to porous heterogeneities in dual-porosity rocks. Their formulations include a homogeneous porous host material with ellipsoidal and spheroidal inclusions representing vrugs and cracks in the rock, respectively. These ideas could be followed to include heterogeneities that are representative of macropores. For instance, it is already well established that thermal transport properties are critically dependent on the shape (e.g., spherical and cylindrical inclusions) of the grains and their spatial arrangement (De Vries, 1963). Similarly, Berryman (1995) reviews EMT approaches for modeling the elastic moduli of multiphase materials and illustrates how different inclusions shapes can be accounted for in such formulations. One could also explore advanced simulation methodologies. For example, Rubino et al. (2016) simulate the effective elastic properties of an experimental volume by considering attenuation and dispersion of seismic waves introduced by wave-induced fluid flow (see Pride, 2005) due to mesoscopic (larger than pore scale and smaller than the wavelength) heterogeneities within a porous media. By introducing heterogeneities that are representative of the structural form of a soil (e.g., a hole created by an earthworm), such methodologies could provide insights about the sensitivity of elastic properties (and seismic velocities) to structural form and water saturation.

4.2. Combination of Geophysical Measurement Methods

Geophysical methods differ in their sensitivity to different soil physical properties, some respond primarily to interfaces (wave-based physics: seismic and GPR reflection methods), whereas others to bulk properties (diffusion-based physics: DC-resistivity, and EMI). It is thus conceivable that the combination of geophysical methods may have a synergistic influence on the inferences especially as related to an elusive trait such as soil structure. We discriminate below between multimethod approaches that (i) provide information about soil structure properties and (ii) delineate zones with different soil properties.

The first category is motivated by the fact that a single geophysical property is often insufficient to draw reliable conclusions about soil structure. For instance, the IP method provides information about the conductive and capacitive properties of a soil. Compared to DC-resistivity data that only sense conductive properties, this makes it much easier to attribute responses to salinity, clay, or water content variations (Revil et al., 2017). Similarly, estimations of water content using GPR may be useful to constrain the interpretation not only of electrical resistivity sections (or maps) derived from DC-resistivity (or EMI) but also of seismic velocities.

The second category is motivated by the fact that certain methods (e.g., seismics and GPR) have superior resolution, while the physical property of interest is primarily sensed by a lower-resolution method. As an example, one could improve the information content of DC-resistivity data by constraining its interpretation to soil layers derived from GPR or seismic data (see Doetsch et al., 2010, for a hydrogeological example). This could help to better constrain hydrological dynamics in time-lapse studies (discussed in section 4.4). Another option would be to combine GPR reflectivity patterns and seismic bulk properties to differentiate between loose and compacted soil zones. Similar combinations of GPR and seismics might help to localize large root volumes in forest soils.

Inversion (e.g., Menke, 2012) is needed to translate geophysical data into geophysical properties. The resolution and robustness of geophysical inversion results are often improved by considering multiple data sets or information gained from other types of geophysical data (Linde & Doetsch, 2016). A more advanced approach would, therefore, be to jointly invert the data by directly targeting properties of interest. For instance, both DC-resistivity and GPR data are highly sensitive to porosity (equations (1) and (5)). This suggests that joint inversion schemes combined with appropriate pedophysical parametrizations of soil (in order to target, say, macroporosity rather than total porosity) would be possible. Joint inversion implementations remain challenging since data errors and sensitivities are different for different methods.

Similarly, joint inversion could be used to (i) better understand field-scale petrophysical relationships (e.g., Linde & Doetsch, 2010) and to (ii) enable clustering into zones of distinct soil structure. One way to achieve this is joint inversion using structural constraints in which no petrophysical relationship is imposed (Doetsch et al., 2010; Gallardo & Meju, 2003). Gallardo and Meju (2003) pioneered such an approach by jointly inverting DC-resistivity data and seismic travel-time data. By observing the resulting cross plot of electrical resistivity and *P* wave velocity, they recognized segregation of the scattered points into different zones with different trends suggesting local field-scale petrophysical relationships. Doetsch et al. (2010) used such plots as input to clustering algorithms that provided a zonation of the subsurface.

4.3. Survey Design and Spatial Scaling of Soil Structural Features

Geophysics may offer extensive spatial coverage that provides a more integrative view of the subsurface than what can be obtained by sparse soil sampling or deployment of point sensors. In general, the support volumes of geophysical data and corresponding property estimates obtained by inversion are often much larger than the Representative Elementary Volume scale at which petrophysical models are defined. Any attempts to describe processes or downscale geophysical information to finer spatial scales should consider a certain level of stochasticity. Geostatistical theory offers means to explicitly account for spatial averaging and support volume (Kyriakidis, 2004). Ignoring the scale disparity between geophysical estimates and variables of interest is known to bias results (Day-Lewis et al., 2005), and it might lead to false conclusions. Hence, once a clearer understanding on the signatures of soil structure in geophysical signals is developed, geophysics could help bridging the spatial gap between the very small support volume offered by spatially undersampled subsurface soil sensors and the larger scales that are relevant for land management and climate modeling. Such methods are not yet developed, and refinement of geophysical methods and interpretation would be required to allow for large-scale investigations at high resolution (also with depth).

The main challenge with high-resolution geophysical imaging of soils over large scales is that the horizontal scales are much larger (kilometric or more) than the vertical scale (decimetric to metric) of interest. Exhaustive representation of such landscapes at the resolution relevant to soil structure studies would require the acquisition and processing of massive amounts of measurements (both are time and cost extensive). Experimental setups would also need to be adapted to enable accurate imaging of features at scales of decimeters or less (e.g., account for electrode shape, Rücker & Günther, 2011; and use nonstandard high-frequency seismic sources, Ryden & Lowe, 2004). In addition, noise becomes an issue at these scales, at which electrode contacts exact electrode locations (and their variation over time), and microtopography may introduce considerable errors and mislead interpretations. Drone-based geophysics or instruments that are towed or attached to agricultural machinery (or autonomous vehicles) are likely to become increasingly available and alleviate some of challenges of data acquisition (e.g., http://vulcanuav.com/; https://gamaya.com/; Rapstine et al., 2017; the processing remains a challenge). Other surrogate variables may help indicating presence and influence of soil structure from rapid infiltration or no runoff (local gravity methods); manipulative experiments with large scale salt application to monitor rate of removal and potential pathways; and use of near-surface seismic surveys to establish baseline compaction levels in agricultural regions. To address the data processing and information extraction burden, we envision the use of hierarchical approaches that retain essential soil structural features based on the intended application. The potential of information compression (similar to image compression) could be useful in communicating such large soil structural data sets. We suggest further that low-resolution mapping with EMI or satellite-based remote sensing (that only senses land-surface properties) can be used both to guide the locations of such detailed surveys and to interpolate in-between (Dafflon et al., 2017). Versatile geostatistical techniques that can handle incomplete sampling and different support-scales need to be further developed and demonstrated for this type of problem (Straubhaar et al., 2016). We also expect that recent advances in deep learning can help to address downscaling using image superresolution methods (P. Lu et al., 2018; Shen, 2018).

4.4. Soil Dynamic Responses and Hydrogeophysical Modeling

Repeated geophysical surveys or semipermanent geophysical monitoring make it possible to deduce influences of soil structure indirectly via its impact on various soil dynamic processes. Examples include the activation of macropores (Moysey & Liu, 2012), rapid drainage, salinity changes due to root water uptake, and abrupt changes in seismic or GPR signals following compaction. Given their larger spatial footprint, time-lapse geophysics may offer insights that are impossible when using individual point sensors. In the vein of hydrogeophysics (Binley et al., 2015), we propose to capitalize on the combination of spatial coverage and dynamic monitoring to differentiate geophysical responses related to the presence or absence of certain structural

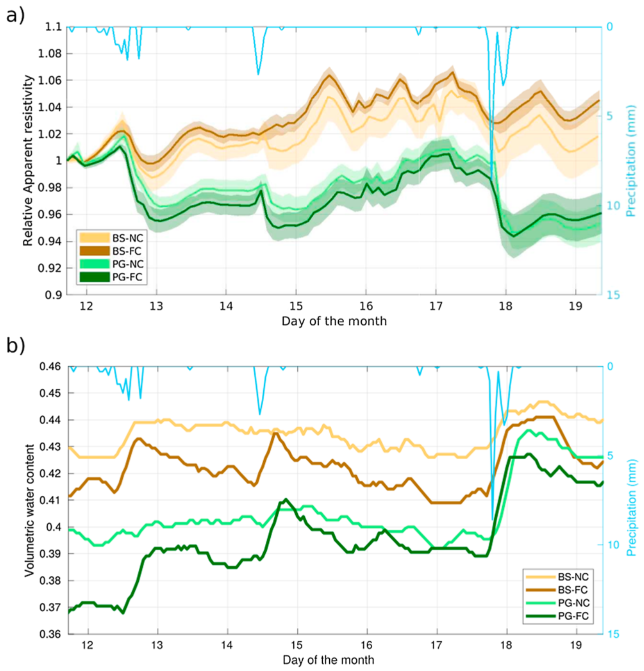

Figure 9. (a) Relative time-lapse changes in electrical resistivity of the first level of the Wenner array for bare soil compacted (BS-FC), bare soil noncompacted (BS-NC), permanent grass compacted (PG-FC), and permanent grass noncompacted (PG-NC). (b) Volumetric water content estimated from Time Domain Reflectometry measurements at 20-cm depth, located in the same plots as in (a).

features. We highlight that the relevant time scales might be years, even decades (aggregation, soil structure formation, and regeneration) and that geophysical monitoring can be achieved over such time scales. Having the possibility of gathering dynamic responses enhances the value of geophysical methods as tools for soil structure characterization. Current instrumentation may not be well suited for such challenging task, the creation of new autonomous machines specifically designed to monitor large areas over long time periods may substantially improve monitoring capabilities.

As an example, we present a week-long time-lapse DC-resistivity monitoring experiment carried out in September 2017 at the Soil Structure Observatory in Zürich, Switzerland (see Keller et al., 2017, for experimental details and Appendix A). Figure 9 presents relative variations (and associated standard deviations) of apparent resistivities (Figure 9a) corresponding to electrodes separated by 50 cm (so-called Wenner array) and volumetric water content estimated using Topp's equation from measurements of TDR probes at a depth of 20 cm (Figure 9b). The data were acquired in four subplots with different characteristics: noncompacted bare soil (yellow), compacted bare soil (brown), noncompacted soil under grassland (mixture of grass and legumes, light green), and compacted soil under grassland (mixture of grass and legumes, dark green). These results obtained 3.5 years after the compaction event can be interpreted by qualitative arguments. The grass-covered zones respond much more to the rainfall on 17 September than the zones with bare soil. This confirms that the bare soil is nearly water-saturated before this precipitation event (see Figure 9b). In contrast, the grass-covered zones present a lower initial water saturation/content (more evapotranspiration and faster downward percolation). Additionally, note how Figure 9 resembles the results obtained by Moysey and Liu (2012; Figure 8); normalized apparent resistivity decreases as the biopores created by the grass and legume roots gradually become active in response to the rainfall. This is not evident for the bare soil, where root-introduced biopores are not developing. The only clear discriminator between the subplots is their structural form, and yet the DC-resistivity data presented here and in Figure 4b could capture aspects related to soil structure (compacted vs. noncompacted) and management (bare vs. grass-covered soil). These observations were obtained from one geophysical method, and interpretations were made with the help of TDR data and based on prior knowledge about the characteristics of the controlled experiment. This enhances the value of the combination of geophysical methods with other sources of information. Additionally, we expect that the combination of methods (see section 4.2) can guide interpretations at locations with more limited prior knowledge about the soil and its history. This example emphasizes the lack of quantitative tools to interpret the observations and reinforces the need for developments described in this section.

The anecdotal applications described above clearly point to the need for methods that fuse spatial and temporal geophysical information to derive quantitative metrics related to soil structure. Such methods could be fashioned after large-scale hydrological and climate models (Looy et al., 2017) or, in the context of geophysics, could use coupled hydrogeophysical inversion to infer subsurface properties (Linde & Doetsch, 2016). In such an approach, the inverse problem is parameterized in terms of hydrological properties. The mismatch between the observed geophysical data and those predicted from the hydrological forward response (using a petrophysical relationship and a geophysical forward model) is used to iteratively infer a set of hydrological parameters that honor the data (Jougnot et al., 2015; Kowalsky et al., 2004; Kuhl et al., 2018; Tran et al., 2016). Thus, the dynamic geophysical response to soil state is used to infer soil properties of interest. Such coupled inversion methodologies have the potential of providing valuable information about the evolution of soil structure and, in the case of degradation, the major aspects influencing its recovery. A further step would be to shift emphasis from parameter estimation toward a framework of formal testing of competing hypotheses of the mechanisms governing soil structure and its development. This can be addressed by model

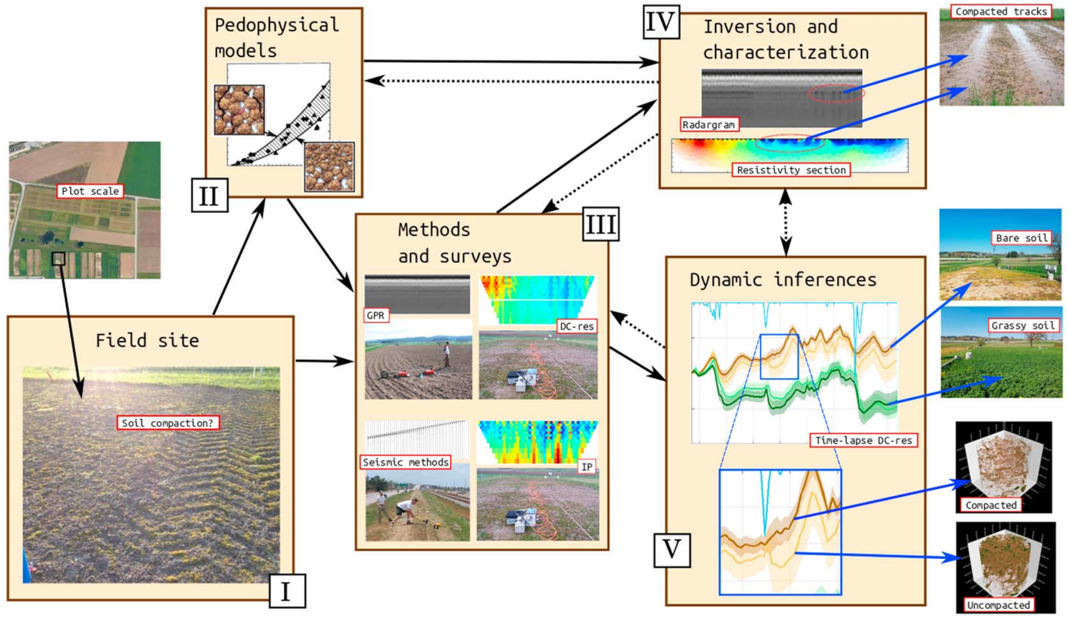

Figure 10. Sketch of how to use geophysical methods for soil structure characterization: a soil compaction example. Black solid lines represent the information flow, black-dashed lines represent feedbacks between activities, and blue lines represent inferences of soil structure. (I) The example shows how geophysical methods can be used to infer information related to soil compaction in an agricultural field. (II) Knowledge about the pedophysical links between geophysical properties and soil structural properties is used to guide the selection of methods and help interpretation. (III) DC-res, IP, GPR, and seismic methods are suitable for obtaining information about soil structure. (IV) The spatial distribution of compaction is captured by the inverted and/or processed geophysical data: track compacted zones produce more conductive regions (DC-res) and enhancement of GPR early signals (see blue arrows). (V) Time-lapse changes in apparent resistivity due to the soil hydrological response to precipitation help to discriminate between soil structures: the grassy soil is highly responsive to rain relative to bare soil, and the uncompacted bare soil is more responsive to rain than the compacted bare soil (blue arrows). GPR = Ground-Penetrating Radar; DC-res = Direct Current-resistivity; IP = Induced Polarization.

selection techniques that use geophysical data to discriminate among multiple competing conceptual soil models (Brunetti et al., 2017; Linde, 2014).

4.5. Summary

This section presents an overview of geophysical methods and models (petrophysical and inversion strategies) for soil structure characterization that have not been yet fully developed, yet they deem to hold most promise for progress in this area. In particular, the harnessing of combination of methods with different sensitivities to soil structure attributes and placing these on a soil structure-aware modeling framework that capitalizes on spatial or temporal signatures (i.e., system dynamics). As a specific example, Figure 10 shows how DC-resistivity and GPR methods can be used to characterize the evolution of different soil structures in relation to soil compaction and soil management (see also Appendix A). The experimental field presents different compaction patterns as well as different soil postcompaction treatments. The characterization of soil surfaces using GPR and DC-resistivity arrays at sufficient spatial resolutions has been shown to identify zones affected by soil compaction. Shortly after the compaction event, these zones would appear as high electrical conductivity (e.g., due to higher saturation levels and volumetric clay content) or increase the amplitude of GPR signals in the radargram at the same spatial locations. Such spatial characterization is not strongly dependent on the dynamics of soil processes, and we thus term this class of characterization *static*. In contrast, repetitive measurements in time could be acquired (e.g., using time-lapse DC-resistivity) with the explicit objective of capturing dynamic (and relatively rapid) changes in the hydrogeophysical state of the soil system. These changes could be fed into a modeling framework that explicitly considers soil structure effects on soil dynamic processes, and thus, we may use these geophysical observations (and inversion framework) to distinguish different integrative signatures of soil structure. The success of new geophysical applications is also critically dependent on developing new and well-tested pedophysical relations that link soil structure states with geophysical measurements.

5. Conclusions

Soil structure governs a wide range of hydrological and ecological soil functions, but there are currently no satisfying ways to measure it noninvasively and at relevant field scales. We have examined how geophysics, with its ability to image large spatial and temporal domains, can be used to obtain insights about soil structure. Many geophysical properties respond to soil structure, but there is a lack of petrophysical (pedophysical) models that relate them to soil structure attributes (e.g., macroporosity and its connectivity). We highlight the need to reduce interpretation ambiguity and increase image resolution by combining multiple geophysical data types. We suggest that indirect inference of soil structure by relating the geophysical time-lapse response to the dynamics of state variables together with appropriate hydromechanical and biological modeling offers multiple possibilities for quantitative assessments of soil structure. Given the many factors that may influence geophysical responses, we suggest that geophysical methodologies for soil structure characterization are first developed and tested at well-instrumented field sites.

Appendix A: Soil Structure Observatories

Field-scale research observatories help to advance science as they concentrate research resources to specific areas where researchers have access to long-term time series and other supporting data sources. Multiple large-scale initiatives related to critical zone science (http://criticalzone.org/national/), climate change (http://www.tereno.net/overview-en?set_language=en), and hydrogeology (http://hplus.ore.fr/en/) have been developed. Similar observatories that are dedicated to soil structure would facilitate the development of geophysical approaches for soil structure characterization. The research opportunities highlighted above would all be facilitated by working on sites where:

1. The experimental design can be controlled (or at least having detailed knowledge).
2. There is easy access to multiple types of data to improve interpretations and, if possible, guide survey design.
3. There is access to laboratory measurements to better understand site pedophysics.
4. It is possible to perform long-term monitoring of soil structure evolution without removing the sensors.

The Soil Structure Observatory (Keller et al., 2017) established in the vicinity of Zürich, Switzerland, seeks to study the evolution of soil structure by measuring different soil properties with point probe measurements, soil sampling, weather measurements, and geophysical techniques. Figures 4 and 9 represent preliminary analysis of DC-resistivity insights related to soil structure, and Figure 5 presents the GPR patterns due to compaction. Clearly, different climates and textures will influence the development of soil structure and responses of soils to deformations, the time needed for them to regenerate after a plastic deformation, and the relative relevance of the different processes that act on them. Consequently, additional Soil Structure Observatories that are representative for other soil types and climatic conditions would be most useful.

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