**Predicting the relative density from on-the-go horizontal
penetrometer measurements at some arable top soils in Northern Switzerland**

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ABSTRACT

This study sought to develop empirical models to predict soil relative density (ρ_{rel}) from measurements of horizontal penetrometer resistance (PR) and soil water content (θ_g) in a wide range of soil textures. This permits the comparison of the state of soil compactness in different soil textures. It was hypothesised that model coefficients would be texture-dependent when soil compactness was expressed as bulk density (ρ_d) and that a model with constant coefficients could be obtained when soil compactness was expressed in terms of ρ_{rel} (obtained as the ratio of ρ_d to reference bulk density (ρ_{ref})). Field measurements were conducted in 2014 using a horizontal penetrometer at 0.25 m depth in 10 fields in Switzerland with a wide range of soil textures covering sandy loam, silt loam, loam, clay loam and clay (clay concentration, (CC) = 153-585 g kg⁻¹ and organic matter concentration, (OM) = 9-168 g kg⁻¹). At selected locations along the penetrometer measurement transects, cylindrical soil cores were sampled for determination of soil texture, OM, θ_g and ρ_d . Soil water potential and effective stress (σ') were also estimated for each location. Standard Proctor tests were performed on eight soils with variable textures. Proctor density was well described as a function of CC and OM (R^2_{adj} = 0.97, RMSE= 0.046 Mg m⁻³) and was used as reference density to obtain ρ_{rel} . From this we developed a model for prediction of ρ_{rel} from PR and σ' that allows comparisons between soils without changes in model coefficients. However, σ' cannot be obtained from on-the-go measurements and the model is therefore of limited value for soil compaction mapping. A model for estimating ρ_{rel} from PR and θ_g yielded satisfactory predictions (R^2_{adj} = 0.66, RMSE= 3.3%), although θ_g is a texture-dependent measure of soil water that cannot be compared across soils. Moreover, ρ_d was well predicted from PR and θ_g (R^2_{adj} = 0.93, RMSE= 0.05 Mg m⁻³), possibly because all our measurements were carried out at similar soil water potential, which implies that θ_g carries soil

66 textural information. Future research should test the proposed equations for a wide range of soil
67 water potential values. The findings presented can be of use in developing measurement systems
68 for mapping soil compactness that combine the proposed prediction functions with horizontal
69 penetrometer and water content sensor systems.

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71 **Keywords:** Soil sensor, Within-field variability, Relative density, Mechanical resistance,

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73 Prediction models

1. Introduction

Soil compaction (i.e. reduction of soil porosity) caused by agricultural field traffic is one of the main threats to sustaining soil quality in Europe (COM, 2006) and worldwide (Oldeman et al. 1991; Farrakh Nawaz et al., 2013). A range of important ecological functions is affected when soil is compacted (e.g. van Ouwerkerk and Soane, 1995; Alaoui et al., 2011). For example, compaction reduces saturated hydraulic conductivity and thus triggers surface runoff and soil erosion by water and also reduces soil aeration and increases mechanical impedance to plant roots and the energy required for mechanical tillage. Consequently, soil compaction is one of the causes of a number of environmental and agronomic problems (flooding, erosion, greenhouse gas emissions, leaching of chemicals to water bodies, crop yield losses) resulting in significant economic damage to society and agriculture (Taylor, 1992; Hamza and Anderson, 2005).

Mapping the state of soil compaction across agricultural fields and landscapes would provide (i) better estimates of the extent of land degradation by compaction (severity of compaction and area affected), and (ii) information on soil traffic management and land use impacts on soil compaction, which in turn would help in the development and refinement of sustainable management guidelines. Moreover, the identification and delineation of areas with a similar degree of soil compactness within arable fields could help to explain within-field crop yield patterns (e.g. Keller et al., 2012) and allow farmers (i) to apply site-specific soil management (e.g. site-specific tillage) (Raper, 1999; Khalilian et al., 2002, Raper et al., 2005), and (ii) to achieve better timeliness in field operations.

A suitable sensor for on-the-go measurement of the state of soil compaction must permit measurements in a soil profile to enable site-specific delineation of the thickness and depth of compacted layers, e.g. for precision (variable-depth) tillage applications. This cannot be achieved

(as yet) by non-intrusive sensors such as remote sensing or surface-contact sensors, e.g. seismic surface wave sensors (Donohue et al., 2013) or electrical conductivity sensors (see Naderi-Boldaji et al., 2014). Moreover, soil density cannot be measured directly with on-the-go methods. Most suitable on-the-go devices for such applications seem to be soil strength sensors (Hemmat and Adamchuk, 2008), typically in the form of horizontal penetrometers that can be pulled through the soil by a tractor. Horizontal penetrometers operate at a specific soil depth, but multi-probe tools have been designed that measure PR simultaneously at various depths (e.g. Chukwu and Bowers, 2005; Chung et al., 2006). Moreover, horizontal penetrometers could be integrated on farm implements (e.g. in front of a seed drill), so that no extra field operations are needed when mapping soil compaction.

PR is an attractive measurement because it is relatively simple, fast and cheap, can be carried out on-the-go, thus yielding spatial information, and is sensitive to soil bulk density, which makes it an excellent measure for assessing soil compaction. However, PR is strongly influenced by soil water content and soil texture, and therefore water content and texture have to be measured simultaneously with PR. This can be achieved by a ‘sensor fusion system’ as demonstrated by e.g. Naderi-Boldaji et al. (2013a).

Relationships between vertical penetrometer resistance (Q) (i.e. cone index) and soil characteristics have been the subject of numerous studies. A variety of empirical functions have been developed that express Q as a function of bulk density (ρ_d) and some measure of soil water, either in the form of water content, water potential or effective stress (for an overview, see Dexter et al. (2007) and Vaz et al. (2011)). It is worth mentioning that Q and PR are strongly correlated but are not identical, due to their different direction of motion (i.e. vertical compared with horizontal) and that the correlation between Q and PR is affected by cone size and working

depth (see Hemmat and Adamchuk, 2008). However, both Q and PR are similarly affected by soil properties (texture, density, water content) and therefore similar models can be used to describe PR and Q as a function of water content and bulk density, although the model coefficients may differ. Some of these functions are purely empirical, while others may be classified as ‘semi-physical’ (equations derived based on physical considerations, but where the coefficients are empirically derived). The equations usually work very well for a particular soil. When a range of soils is considered, the typical approach is to develop empirical relationships between the model coefficients and soil basic properties (soil texture, organic matter concentration) (e.g. Silva and Kay, 1997). Prediction of penetrometer resistance from bulk density and soil water content is a useful tool in studies on root growth, draught requirement of tillage implements or trafficability (Dexter et al., 2007).

In the context of the present study, i.e. soil compaction mapping, the focus is on predicting soil density from penetrometer measurements (in contrast to the work cited above, which predicts penetrometer resistance from bulk density). Because penetrometer resistance is strongly influenced by soil water, such a model will have the general form:

$$\rho_d = f(Q, m) \quad (1)$$

where m is an expression of soil water content. An example of such a model for prediction of ρ_d was developed by Mouazen et al. (2003) and tested by Mouazen and Ramon (2006) on a 2.3 ha field using on-the-go measurements of draught force of a soil cutting tool (i.e. a chisel) and water content. Similarly to the models for prediction of Q , a model for estimation of ρ_d can be expected to perform well for any particular soil. For example, the field investigated by Mouazen and

Ramon (2006) varied in texture within a relatively narrow range (clay concentration between 75 and 92 g kg⁻¹). However, such a model is expected to perform less well when different soils are compared. As with the prediction models for Q , a possible approach is to combine Eq. (1) with a number of empirical regression equations for estimation of model coefficients from easily-available soil properties. Mouazen and Ramon (2009) and Quraishi and Mouazen (2013) added empirical correction factors that could be predicted from soil texture to the model developed by Mouazen and Ramon (2006). In this study, we assumed that the density in Eq. (1) could be expressed in terms of relative density. The use of relative density is intended to normalise bulk density in such a way that a certain relative density describes the same state of compactness in any soil, i.e. independently of soil texture. Hence, relative density can be used for comparisons across soil types. In contrast, values of absolute bulk density may indicate a dense state in one soil, but a loose state in another soil. Consequently, optimum and critical limits of bulk density for crop growth depend strongly on soil texture (e.g. Håkansson, 1990). An expression of relative density (ρ_{rel}) that is widely used in agricultural research is the degree of compactness (DC) devised by Håkansson (1990). It is simply expressed, typically in %, as the ratio of ρ_d to a reference bulk density (ρ_{ref}):

$$\rho_{rel} = \frac{\rho_d}{\rho_{ref}} \times 100 \quad (2)$$

Different methods have been proposed for obtaining ρ_{ref} , such as uniaxial compression or Proctor tests (for an overview, see Reichert et al. 2009). Håkansson (1990) specially developed a long-term (about one week) drained uniaxial compression test on large loose soil samples at 200 kPa to obtain reference density. In this research, however, we used the Proctor test, because the

resulting Proctor density is independent of initial soil structure and initial soil water content, and because the Proctor test is less laborious than the original [Håkansson \(1990\)](#) reference density test ([Naderi-Boldaji and Keller, 2016](#)). Furthermore, the Proctor apparatus is more widely available in laboratories than other equipment for compaction tests.

The aim of this study was to develop a physically sound empirical model for prediction of relative density from PR and water content, based on on-the-go measurements of PR with a horizontal penetrometer. The PR was measured on arable fields in Switzerland that cover a wide range of soil textures. In a first step, we developed an equation for prediction of bulk density from PR and water content. We hypothesised that such a model is soil texture-dependent (i.e. the model coefficients are functions of texture), while the use of relative density would remove the texture dependency (i.e. the coefficients would be constants and the same equation could be used for comparison of different soils). Furthermore, we examined the use of different expressions of soil water (water content, relative water content and effective stress) in the prediction models.

2. Materials and methods

2.1. Field measurements

Field measurements were conducted in summer 2014 in 10 arable fields (F1-F10) with different soil textures in Northern Switzerland (Table 1). Most of the fields were located at the Agroscope research station in Zürich, one field was close to Solothurn and another was near Basel. In addition, a data set (48 measurement points) from a previous study ([Naderi-Boldaji et al., 2013a](#)) conducted near Bern, Switzerland, was included (F7, cf. Table 1). The fields are annually mouldboard-ploughed to a depth of 0.2 m. To capture a range of variations in water

content and bulk density, in some fields (i.e. F1, F3 and F4), where available soil maps showed a narrow range of texture variability, transect measurements were conducted in a limited area of the fields but on different occasions (i.e. at different water contents), while for the rest of the fields with higher texture variability, measurements were made within different areas of the field on only one occasion (but not on the same date for all the fields). Therefore, for fields F1, F3 and F4, only the soil at one location was analysed for texture and OM as a representative soil of the area, while for the other fields at least one sample from each transect was analysed for texture and OM. For the 48 locations of the 2013 study (Naderi-Boldaji et al., 2013a), the soil texture and OM were analysed for all points.

2.2. Horizontal penetrometer

A single-tip horizontal penetrometer previously developed by Naderi-Boldaji et al. (2012) was used for field measurements of soil penetration resistance. The horizontal penetrometer (Fig. 1) consists of 1) a frame for fully-mounting on a tractor and gauge wheels for adjusting the working/measurement depth, 2) a tine to cut through the soil, 3) a horizontal cone and shaft as the sensing tip and 4) a strain-gauged load cell and an interfaced data acquisition system for measuring the soil resistance to the penetrometer. Here, a 30 mm diameter cone with an apex angle of 45° was connected to a 5000-N load cell (Bongshin Model DBBP, Bongshin Load Cell Co., Ltd., Korea) with a nominal sensitivity of 3mV V⁻¹ behind the tine that measured PR directly transmitted from the conical tip. The output of the load cell was logged by a commercial IOtech data logger (IOtech, Personal Daq/56™ Series, USA) with a sampling rate of 32 Hz. Additional weights were placed on the penetrometer frame to prevent up-lift

during measurement. The vertical position of the tine was adjusted by the top-link of the tractor's three-point hitch.

For the field measurements, the working depth of the horizontal penetrometer was 0.25 m and the forward speed was controlled at 0.5 m s^{-1} by a proper combination of gear ratio and engine speed. The working depth was set so that the sensing tip of the horizontal penetrometer was located below the critical depth of the penetrometer tine where soil is not already disturbed by the tine (Naderi-Boldaji et al., 2013b). The average length of transects in different fields (other than F7) was about 30 m. Longer transects were not measured in order to avoid soil texture variations. Immediately after PR measurements, the PR signal of each transect was analysed/synchronised with respect to the transect length in order to select the locations for sampling. To capture as wide a range of PR as possible, the locations were selected at local PR minima and maxima. Typically, 2-5 locations were selected per transect. Fig. 2 shows a typical variation in PR along a transect in F1 and the locations selected for core sampling. Two undisturbed cylindrical soil cores (100 mm diameter and 60 mm height) were then sampled at 0.25 m depth (working depth of the horizontal penetrometer) at each location, for determination of water content and bulk density by the oven-drying method. The raw PR signals were filtered with a Butterworth filter with a frequency of 0.1 Hz to remove any sharp (spiky) fluctuations and subsequently averaged for 1 m ($\pm 0.5 \text{ m}$) at the locations of the core samples. In total, 92 measurement locations were selected on the 10 fields. From eight fields (see Table 1), about 15 kg of disturbed soil were taken for determination of Proctor reference density. For the F7 field (measured in 2012), Proctor tests were made for two zones, one (F7-HC) with high clay concentration (261 g kg^{-1}) and one (F7-LC) with low clay concentration (167 g kg^{-1}) (Table 1).

2.3. Laboratory measurements

Soil particle size distribution was analysed using the pipette method, and organic matter concentration was calculated from the loss on ignition (Schulte and Hopkins, 1996). Particle density was measured using a pycnometer. Standard Proctor tests (ASTM Standard D698, 2007) were carried out for eight different soil textures (Table 1) to estimate the maximum density, which was used in this study as the reference density (Eq. 2). The aim was to find a relationship for prediction of Proctor density as a function of soil texture that could be applied for the three other soils for which the Proctor test could not be performed, and also for future studies.

For the Proctor tests, the disturbed soils were sieved at 8 mm and dried at 105 °C for 48 h to ensure zero water content. Then, soils with differing water content were prepared by spraying on a calculated mass of water and gently mixing the soil to achieve uniform water distribution. The prepared samples were tightly covered with plastic sheets and stored in a cool room for 24 h before the Proctor tests to allow for water redistribution and ensure a uniform water content. The tests were carried out using an electrical compaction machine (mould volume: 942 cm³, mould inner diameter: 105 mm, hammer weight: 24.5 N, falling height: 305 mm, with 25 blows on each layer) with a compaction energy of 0.6 MJ m⁻³.

2.4. Estimations of water retention characteristics and effective stress

The water potential (h) for each sampling location was estimated on the basis of the van Genuchten (1980) equation:

$$\theta_g = (\theta_{sat} - \theta_{res})[1 + (\alpha h)^n]^{-m} + \theta_{res} \quad (3)$$

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264 where θ_g (g g⁻¹) is the gravimetric water content at suction h , θ_{sat} and θ_{res} are the saturated and
 265 residual water content, respectively, h (hPa) is the pore water suction (i.e. the negative of the
 266 matric water potential), α (hPa⁻¹) is a reciprocal suction that is characteristic for the soil, and m
 267 and n are dimensionless variables that describe the shape of the curve. Solving Eq. (3) for h
 268 yields (see also [Dexter et al., 2007](#)):

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$$h(\chi) = \frac{1}{\alpha} [\chi^{-\frac{1}{m}} - 1]^{\frac{1}{n}} \quad (4)$$

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272 where χ is the degree of saturation given as:

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$$\chi = \frac{(\theta_g - \theta_{res})}{(\theta_{sat} - \theta_{res})} \quad (5)$$

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276 We applied the [Mualem \(1976\)](#) constraint:

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$$m = 1 - \frac{1}{n} \quad (6)$$

279

280 The parameters in Eqs. (3-6), i.e. α , n , θ_{res} and θ_{sat} , were estimated by means of the pedo-
 281 transfer functions by [Wösten et al. \(1999\)](#). The effective stress (σ') was then calculated from
 282 ([Towner and Childs, 1972](#)):

283

$$\sigma' = \chi h \quad (7)$$

Eq. (7) ignores the contribution to σ' from the surface tension in water menisci between soil particles. However, it has been shown that this component of the effective stress can be neglected for $\chi > 0.4$ (see Dexter et al., 2007), which applies in our case.

We also investigated the interaction between water content and bulk density for different soil textural classes. For this, we calculated the water retention curve (WRC) for the 12 FAO/USDA textural classes using the van Genuchten (1980) equation (Eq. (3)) in combination with the pedo-transfer functions of Wösten et al. (1999). For each class, we used the typical values for texture and OM given in Table 1 of Dexter (2004). Bulk density was applied in the range 0.8-1.8 Mg m⁻³ at intervals of 0.2 Mg m⁻³ for each textural class, while all the other parameters (i.e. texture and OM) were kept constant. Note that the bulk density range tested here may not be realistic for all texture classes, but the wide range in bulk densities is helpful for the purposes of illustration.

2.5. Development of prediction equations

Measured values of bulk density and relative density were regressed against measured PR and θ_g and estimated σ' . In order to develop the prediction models, the stepwise selection method of SAS (SAS Institute INC., Cary, NC, USA) was used. Factors were only included in the models if they were significant at $P < 0.1$.

3. Results and discussions

3.1. Proctor density

The Proctor maximum density (ρ_{ref}) and optimum water content (θ_{opt} ; i.e. θ_g at the maximum density) are presented in Table 2. It can be seen that ρ_{ref} decreased and θ_{opt} increased with increasing OM and clay concentration (CC). We obtained:

$$\rho_{ref} = 1.9 - (0.00342 \times OM) - (0.000529 \times CC) \quad (8)$$

$$R^2_{adj} = 0.96, RMSE = 0.046 \text{ Mg m}^{-3}, P < 0.0001$$

where ρ_{ref} is in Mg m^{-3} and OM and CC are in g kg^{-1} . Fig. 3 depicts the correlation between measured and predicted ρ_{ref} by Eq. 8. The equation was used to estimate ρ_{ref} for the three soils where ρ_{ref} was not measured (F8, F9 and F10, cf. Table 1). Relative density was then calculated for each sampling location using Eq. 2.

3.2. Correlations between soil characteristics and *PR*

The descriptive statistics between soil characteristics and *PR* are given in Table 3 and the correlations in Table 4. The clay concentration ranged between 153 (F7) and 585 g kg^{-1} (F6). One soil (F3) had a high OM concentration (168 g kg^{-1}), while OM ranged from 9 to 56 g kg^{-1} in the other soils. The bulk density (ρ_d) was in the range 0.73-1.7 Mg m^{-3} and the gravimetric water

content (θ_g) ranged from 0.17 to 0.82 g g⁻¹. Thus, the soils covered a wide variation in soil texture and OM, as well as in ρ_d and θ_g . Measured **PR** was in the range 0.35-1.91 MPa.

PR was negatively correlated with soil water content (θ_g , or volumetric water content (θ_v)) and positively correlated with soil density (ρ_d or ρ_{rel}), as shown in Table 4. It was found that **PR** decreased with increasing CC and OM. This can be explained by strong positive correlations between CC, OM and θ_g . Moreover, ρ_d was negatively correlated with CC and OM, which is in agreement with previous work (e.g. Soane, 1990; Dexter, 2004; Dexter et al., 2005; Keller and Håkansson, 2010).

It is found that ρ_{rel} was less affected by soil texture (CC) and OM than ρ_d , which indicates that ρ_{rel} is more appropriate than ρ_d for comparisons between soils. A very strong correlation was found between θ_g and ρ_d , as discussed in detail in section 3.3.1.

Fig. 4a-c show **PR** as a function of θ_g (Fig. 4a), ρ_d (Fig. 4b) and ρ_{rel} (Fig. 4c) for all the test soils. For better presentation, we pooled our soils with respect to the USDA textural classes. The relationship between **PR** and θ_g was similar for most soils, but the clay soils (F3 and F6) clearly differed from the rest (Fig. 4a). A similar trend was observed for the relationship between **PR** and ρ_d , with the clay soils deviating significantly from the other soil textures (Fig. 4b). However, when soil density was expressed in terms of ρ_{rel} the scatter was smaller and all soils, including the clay soils, fitted on the same linear trend line between **PR** and ρ_{rel} (Fig. 4c).

3.3. Equations for prediction of bulk density from **PR** and water content

We found that bulk density as a function of **PR** and soil water content could be well described by a simple linear equation for our data:

$$\rho_d = a_1 + b_1 PR - c_1 \theta_g - d_1 (PR * \theta_g) \quad (9)$$

In this equation, ρ_d is positively related with PR (as PR increases with increasing ρ_d) and negatively with θ_g (as a denser soil has a lower capacity to store water and vice versa). The interaction term ($PR * \theta_g$) compensates for the effect of soil water content on PR (i.e. PR decreases with increasing θ_g and vice versa). We investigated whether the coefficients of this equation would be different for different soils. Because there were few measurement locations in most fields (cf. n in Table 3), we pooled the data with respect to the FAO/USDA textural classes. We then fitted Eq. 9 to measured data for each textural class separately and regressed its coefficients against CC and OM. The results are summarised in Table 5. **The data support the hypothesis that different soils had different coefficients.** However, the texture dependency was not as strong as we expected, as indicated by the rather low R^2 in Table 5. No impact on b_1 (Eq. 9) of either clay concentration or OM was found. The RMSE (predicted versus measured ρ_d) of the model fitted to different soil textures was highly positively correlated with OM ($R^2=0.8$, data not shown). This may suggest that the predictive power of Eq. 9 decreases with increasing OM. When fitting Eq. 9 to measured data from all soils, we found:

$$\rho_d = 1.69 + 0.25 PR - 0.82 \theta_g - 1.15 (PR * \theta_g) \quad (10)$$

$$R^2_{adj} = 0.93, \text{ RMSE} = 0.05 \text{ Mg m}^{-3}$$

The interaction term ($PR * \theta_g$) was statistically significant and significantly increased R^2_{adj} .

Fig. 5 shows measured ρ_d versus ρ_d estimated using Eq. 10.

Our measurements indicated a non-linear relationships between PR and θ_g (Fig. 4a) and PR and ρ_d (Fig. 4b). Several researchers have suggested describing these relationships with exponential functions (see Vaz et al., 2011), and so we formulated the equation:

$$\rho_d = a_2 e^{(b_2 PR)} e^{(c_2 PR \theta_g)} \quad (11)$$

This equation performed slightly less well than Eq. 10 when fitted to measurements from all soils:

$$\rho_d = 1.55 e^{0.422 PR} e^{-2.057 PR \theta_g} \quad (12)$$

$$R_{adj}^2 = 0.87, \text{RMSE} = 0.07 \text{ Mg m}^{-3}$$

Note that both Eqs. 10 and 12 are physically sound: ρ_d increases with increasing PR, while the impact of θ_g on PR is negative. The non-linear behaviour between PR and θ_g , and PR and ρ_d , could also be described by power-law functions (Hernanz et al., 2000; Vaz et al., 2011), and this would yield:

$$\rho_d = a_3 PR^{b_3} \theta_g^{c_3} \quad (13)$$

However, fitting Eq. 13 to our data yielded a negative coefficient for b_3 , which is not physically meaningful. The reason for this may be the strong correlation between ρ_d and θ_g that we found for our data (Table 4), as shown in Fig. 6a. The regression curve in Fig. 6a is:

$$\rho_d = 2.0 e^{-1.25\theta_g} \quad (14)$$

$$R^2_{adj} = 0.95, \text{ RMSE} = 0.05 \text{ Mg m}^{-3}$$

The R^2 and RMSE of Eq. (14) were almost similar to those of Eq. (10). This implies that ρ_d could simply be predicted from θ_g . This observation prompted closer investigation of the relationship between ρ_d and θ_g .

3.3.1. Relationship between bulk density and water content

We analysed the relationship between θ_g and ρ_d at a range of water potentials for various soil textures. For this, we calculated the water retention curve (WRC) for the 12 FAO/USDA textural classes as described in Section 2.4. Fig. 6b illustrates the calculated relationships between ρ_d and θ_g for the different soil textural classes at a water potential of 100 hPa, i.e. at field capacity. A non-linear relationship between ρ_d and θ_g was found, which could be well described by an exponential function for each soil texture. A power-law function resulted in an equally good fit (not shown). An exponential regression curve of the data for all textural classes yielded:

$$\rho_d = 2.0 e^{-1.44\theta_g} \quad (15)$$

$$R^2_{adj} = 0.90, \text{ RMSE} = 0.12 \text{ Mg m}^{-3}$$

which is similar to Eq. (14), which was derived from our measurements, but with a larger scatter of the data points (i.e. larger RMSE) due to a wider range of soil textures considered for Eq. 15.

The deviation from this regression line was unsatisfactory only for sandy soils (textural classes: sand, loamy sand, and sandy loam).

Next, we tested how the relationship between ρ_d and θ_g would change when different water potentials were considered. Fig. 6c shows ρ_d as a function of θ_g for a loam soil at water potentials of 25, 100 and 250 hPa. Similar results were obtained for the other soil textural classes (data not shown). The exponential ρ_d - θ_g relationship for a given texture was very strong even when the water potential varied between 25 and 250 hPa (Fig. 6c), where 25 hPa represents a wet soil and 250 hPa a relatively dry subsoil (e.g. 0.25 m as in this study) in the study area (Switzerland, temperate climate with annual average precipitation of >1000 mm). For example, the long-term average soil water potential at 0.35 m depth in a loam soil in the Swiss plateau indicates that the soil is dryer than 250 hPa in only about three months of the year (Gut et al., 2015).

Our research suggested that ρ_d and θ_g are highly correlated and that θ_g is a strong predictor of ρ_d when measurements are made within a certain range of soil water potential, such as in wet soil (e.g. in the range 25-250 hPa as demonstrated here). The correlation between ρ_d and θ_g becomes weaker when a larger variation in water potential is considered (e.g. from wet or dry). Note that the ρ_d versus θ_g relationship could be used to predict ρ_d from measurements of θ_g when the focus is on spatial variability (e.g. within-field variability) and measurements are made at similar water potential. However, it may not be useful when measurements from different dates (e.g. with a large difference in water potential between dates) are compared for a specific location. For such a scenario, the water content would change between measurement occasions, but bulk density would not. Hence, PR as an auxiliary variable is needed to accurately predict soil density, requiring the application of a combined penetrometer equipped with a water content sensor.

3.4. Models for prediction of relative density

The contribution of pore water to soil strength is described by the effective stress (σ'), which has the same physical meaning in all soils (Bishop, 1959; Dexter et al., 2007). Dexter et al. (2007) proposed describing Q as a function of the soil quality index (S) (Dexter, 2004) and σ' , and suggested $1/S$ as a measure of soil compactness. Recently, Naderi-Boldaji and Keller (2016) demonstrated that $1/S$ is highly correlated with the degree of soil compactness (DC). Combining these previous findings, we propose that ρ_{rel} can be described as a function of PR and σ' :

$$\rho_{rel} = \alpha_1 + \beta_1 PR + \gamma_1 \sigma' - \delta_1 (PR \sigma') \quad (16)$$

When we regressed the coefficients in Eq. 16 against soil texture and OM, as described above for Eq. (10), we found no impact of soil texture or OM on any of the coefficients. From our measurements, we found that:

$$\rho_{rel} = 77 + 2.5 PR + 0.05 \sigma' - 0.012 (PR \sigma') \quad (17)$$

$$R^2_{adj} = 0.57, \text{ RMSE} = 3.7\%$$

This model yields acceptable predictions, although R^2 is much lower than for Eqs. (10), (12) and (14). One reason for this lower R^2 could be that σ' had to be estimated by means of pedo-transfer functions (cf. section 2.4). Pedo-transfer functions depict general trends, but may produce inaccurate values for any particular soil. We found a strong positive correlation between PR and σ' but a relatively large scatter (Fig. 7). We expect that measurements of h and χ (Eqs. 4

and 5) would reduce the RMSE and increase the R^2 . Nevertheless, the $R^2_{adj}= 0.57$ for Eq. (17) is larger than the $R^2= 0.38$ presented by [Dexter et al. \(2007\)](#) for their model, which describes Q as a function of $1/S$ and σ' .

Although the use of σ' such as in Eq. 17 is the most physically correct way of describing soil water, σ' cannot be measured on-the-go. Therefore, while Eq. 17 is a promising approach and we plan to test this equation in future research, it may be of limited direct use for on-the-go soil compaction mapping.

An expression of soil water that has similar meaning across soil types can be achieved by relating the actual water content to the water content at the plastic limit, PL. The ratio θ_g/PL is widely used in tillage studies (e.g. [Dexter and Bird, 2001](#)). The plastic limit is relatively easily obtained using the “thread rolling method” (Casagrande method, as described in [ASTM Standard D4318, 2007](#)), and such a test could be used in combination with on-the-go field measurements of PR and θ_g . However, this may not be attractive for a real online solution. In that case, PL could be estimated from soil texture and OM, e.g. by using the equations developed by [Keller and Dexter \(2012\)](#), where information on the texture could be obtained from previously developed soil maps or measured on-the-go together with PR and θ_g (e.g. [Naderi-Boldaji et al., 2013a](#)). When we applied Eq. (13) of [Keller and Dexter \(2012\)](#) to our soils, we obtained:

$$\rho_{rel} = 115.7 + 1.4 PR - 26.5 \left(\frac{\theta_g}{PL} \right) - 1.65 \left(PR * \frac{\theta_g}{PL} \right) \quad (18)$$

$$R^2_{adj}= 0.62, \text{ RMSE}= 3.5\%$$

The performance of this equation was slightly better than that of Eq. 17. However, the coefficients changed slightly when Eq. 18 was fitted separately for the different soil textures. We

found that at a given soil water potential, θ_g/PL was not independent of soil texture. The correlation between soil texture and θ_g/PL was weak at the wet end, but increased when the soil water potential became more negative (data not shown).

Finally, we developed equations for prediction of ρ_{rel} from PR and θ_g . Such an approach obviates the need for soil texture information if the model parameters are texture-independent, a significant advantage for on-the-go measurements since depth-specific online measurement of soil texture is not trivial (e.g. Naderi-Boldaji et al., 2013a). Analogously to Eq. 10, a simple linear model is:

$$\rho_{rel} = \alpha_2 + \beta_2 PR + \gamma_2 \theta_g - \delta_2 (PR \theta_g) \quad (19)$$

We found that the coefficients were only slightly affected (values of $R^2 < 0.15$) by soil texture when separately fitting Eq. 19 to each soil texture class and regressing the coefficients against soil texture and OM. When fitting Eq. 19 to the data from all soils, we obtained:

$$\rho_{rel} = 84.62 + 17.03 PR + 10 \theta_g - 74.25 (PR \theta_g) \quad (20)$$

$$R^2_{adj} = 0.66, RMSE = 3.3\%$$

Predicted versus measured ρ_{rel} using Eq. 20 is plotted in Fig. 8. Although θ_g is a measure of soil water content that cannot be compared between soils (a certain value of θ_g may indicate a wet condition in a sandy soil but a dry condition in a clay soil), ρ_{rel} was predicted with a lower RMSE and a higher R^2 with Eq. 20 than with Eq. 17 or Eq. 18. This is probably because our measurements were performed within a narrow range of soil water potential, which implies that

θ_g carries some texture information. For example, at a water potential of 100 hPa (i.e. field capacity), θ_g was strongly correlated with clay concentration (data not shown). A non-linear model, e.g. using exponential functions similar to Eq. 13, did not improve the predictive power.

4. Conclusions

A texture-independent model was obtained for prediction of ρ_{rel} from PR and effective stress (σ' , which accounts for the effects of soil water). This equation (Eq. 17) permits comparisons between soil types without any change in model coefficients. However, σ' cannot be measured by on-the-go sensors and thus the model may be of little direct use for online applications of soil compaction mapping. ρ_{rel} was well predicted (R^2_{adj} = 0.66, RMSE= 3.3%) from PR and θ_g for our data, although slightly different coefficients were found for different soil textures. Estimation of ρ_{rel} from PR and θ_g was possible without the need for a texture-independent measure of soil water (such as σ'), presumably because our measurements were carried out at similar water potential, which implies that θ_g carries information on soil texture. For similar reasons, we found a very strong correlation between ρ_d and θ_g , indicating that ρ_d could simply be estimated from θ_g when comparisons are made between soils within a narrow range of water potential.

Future research could test the proposed equations over a wider range of soil water content. The models evaluated in this study could be applied together with on-the-go measurements using sensor fusion systems composed of a horizontal penetrometer and a water content sensor for measurements of the state of soil compactness in terms of relative density at high spatial resolution. By using a multi-tip horizontal penetrometer, the equations could be examined at

different working depths to enable on-the-go determination of the thickness and the depth of compacted layers for variable-depth tillage application.

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Table 1

Textural composition of the soils studied (at 0.25 m depth). Sand: sand concentration (0.05-0.2 mm); Silt: silt concentration (0.002-0.05 mm); Clay: clay concentration (<0.002 mm); OM: organic matter concentration

Field no.	Sand	Silt	Clay	OM	Particle density	USDA Soil Class	Proctor test
	(g kg ⁻¹)	(g kg ⁻¹)	(g kg ⁻¹)	(g kg ⁻¹)	(Mg m ⁻³)		
F1	242.6	537.1	220.3	15	2.56	Silt loam	*
F2	330.3	475.0	194.7	19	2.55	Loam	*
F3	161.1	358.2	480.8	168	2.34	Clay	*
F4	430.3	284.8	284.8	31	2.59	Clay loam	*
F5	238.0	571.7	190.2	17	2.64	Silt loam	*
F6	39.0	366.0	595.0	52	2.60	Clay	*
F7-LC	537.4	295.8	166.8	23	2.61	Sandy loam	*
F7-HC	399.2	340.1	260.7	18	2.66	Loam	*
F8	320.4	413.3	266.3	19	-	Loam	-
F9	569.1	239.8	191.1	17	-	Sandy loam	-
F10	434.4	305.2	260.4	40	-	Loam	-

*Indicates the soils tested for Proctor reference density.

Table 2

Proctor density (ρ_{ref}) and water content at maximum density (θ_{opt}) of the eight soils tested

Field no.	ρ_{ref} (Mg m ⁻³)	θ_{opt} (g g ⁻¹)
F1	1.725	0.181
F2	1.769	0.154
F3	1.068	0.215
F4	1.679	0.181
F5	1.656	0.178
F6	1.398	0.231
F7-LC *	1.743	0.170
F7-HC *	1.717	0.173

*LC and HC refer to the low and high clay soils of field 7 (F7).

Table 3

Range of soil characteristics and horizontal penetrometer resistance (PR) in the 10 fields studied.

θ_g : gravimetric water content; ρ_d : dry bulk density; ρ_{rel} : relative density; CC: clay concentration;

OM: organic matter concentration; PR: horizontal penetrometer resistance

Field	n*	θ_g	ρ_d	$\rho_{rel}(\%)$	CC	OM	PR
no.		(g g ⁻¹)	(Mg m ⁻³)		(g kg ⁻¹)	(g kg ⁻¹)	(MPa)
F1	9	0.20-0.26	1.46-1.64	84.4-94.9	220	15	0.44-1.65
F2	4	0.19-0.24	1.47-1.56	85-90.2	190-195	19	0.76-1.04
F3	4	0.64-0.82	0.73-0.87	68.5-80.6	481	168	0.44-0.56
F4	5	0.27-0.31	1.28-1.49	78-90.5	285	31	0.35-0.51
F5	5	0.23-0.26	1.44-1.53	83-90.4	190-247	17-21	0.87-1.11
F6	5	0.33-0.5	1.08-1.37	77-86.3	367-585	29-56	0.55-1.01
F7	48	0.17-0.31	1.27-1.69	77.1-96	153-314	9-39	0.88-1.91
F8	4	0.23-0.25	1.44-1.53	84.7-90.1	243-266	19-21	0.44-0.79
F9	4	0.21-0.31	1.29-1.63	78.1-93.9	191-260	17-31	0.42-0.64
F10	4	0.31-0.36	1.26-1.39	77-85.8	225-260	40-41	0.6-0.8

*Number of observations

Table 4

Pearson coefficient of correlation (r) for the relationship between soil characteristics and horizontal penetrometer resistance (PR) for all fields studied. θ_g : gravimetric water content; θ_v : volumetric water content; ρ_d : dry bulk density; ρ_{rel} : relative density; CC: clay concentration; OM: organic matter concentration; PR: horizontal penetrometer resistance. All correlations significant at $P < 0.0001$

Characteristic	θ_g (g g ⁻¹)	θ_v (cm ³ cm ⁻³)	ρ_d (Mg m ⁻³)	ρ_{rel} (%)	CC (g kg ⁻¹)	OM (g kg ⁻¹)	PR (MPa)
θ_g (g g ⁻¹)	1.000						
θ_v (cm ³ cm ⁻³)	0.94	1.000					
ρ_d (Mg m ⁻³)	-0.95	-0.9	1.000				
ρ_{rel} (%)	-0.73	-0.7	0.88	1.000			
CC (g kg ⁻¹)	0.82	0.85	-0.75	-0.45	1.000		
OM (g kg ⁻¹)	0.95	0.82	-0.9	-0.62	0.71	1.000	
PR (MPa)	-0.55	-0.68	0.58	0.56	-0.49	-0.41	1.000

Table 5

The coefficients in Eq. 9 as functions of soil texture. CC: clay concentration; OM: organic matter concentration (g kg^{-1}); R^2_{adj} : adjusted coefficient of determination

Parameter	Equation	R^2_{adj}
a_1	$a_1 = -0.05 \cdot \text{OM} + 3.21$	0.39
b_1	-	-
c_1	$c_1 = 0.17 \cdot \text{OM} - 6.15$	0.32
d_1	$d_1 = 0.011 \cdot \text{CC} - 2.71$	0.25

Fig. 1. The horizontal penetrometer used in this study. From Naderi-Boldaji et al. (2012).

Fig. 2. A typical variation in horizontal penetrometer resistance (PR) along a transect. The circles show the selected locations for core sampling.

Fig. 3. Measured versus estimated Proctor reference density (ρ_{ref}) using Eq. 8. R^2_{adj} : adjusted coefficient of determination; RMSE: root mean squared error.

Fig. 4. Horizontal penetrometer resistance (PR) as a function of (a) water content (θ_g), (b) bulk density (ρ_d) and (c) relative density (ρ_{rel}) for all soils used in this study.

Fig. 5. Measured versus predicted bulk density (ρ_d) using Eq. 10. R^2_{adj} : adjusted coefficient of determination; RMSE: root mean squared error.

Fig. 6. Relationship between bulk density (ρ_d) and water content (θ_g) for (a) field observations, (b) the 12 FAO/USDA soil textural classes at a water potential of 100 hPa, and (c) a loam soil at water potential of 25, 100 and 250 hPa. sa = sand, si = silt, l = loam, cl = clay. See text for details.

Fig. 7. Horizontal penetrometer resistance (PR) as a function of effective stress (σ') for the soils studied.

Fig. 8. Measured versus predicted relative density (ρ_{rel}) using Eq. 20. R^2_{adj} : adjusted coefficient of determination; RMSE: root mean squared error.

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(a)

(b)

(c)

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Fig. 5. Measured versus predicted bulk density (ρ_d) using Eq. 10. R^2_{adj} : adjusted coefficient of determination; RMSE: root mean squared error.

(a)

(b)

(c)

Fig. 6. Relationship between bulk density (ρ_d) and water content (θ_g) for (a) field observations, (b) the 12 FAO/USDA soil textural classes at a water potential of 100 hPa, and (c) a loam soil at water potential of 25, 100 and 250 hPa. sa = sand, si = silt, l = loam, cl = clay. See text for details.

Fig. 7. Horizontal penetrometer resistance (PR) as a function of effective stress (σ') for the soils studied.

Fig. 8. Measured versus predicted relative density (ρ_{rel}) using Eq. 20. R^2_{adj} : adjusted coefficient of determination; RMSE: root mean squared error.