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# Decoupling of topsoil and subsoil controls on organic matter dynamics in the Swiss Alps

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ABSTRACT

Our understanding of mechanisms governing soil organic matter (OM) stability is evolving. It is gradually becoming accepted that soil OM stability is not primarily regulated by the molecular structure of plant inputs, but instead by the biotic and abiotic properties of the edaphic environment. Moreover, several experimental studies conducted in artificial systems have suggested that mechanisms regulating OM stability may differ with depth in the soil profile. Up to now however, there is very limited field-scale evidence regarding the hierarchy of controls on soil OM dynamics and their changes with soil depth.

In this study, we take advantage of the high heterogeneity of ecological conditions occurring in the alpine belt to identify the major determinants of OM dynamics and how their significance varies with depth in the soil profile. Aboveground litter, mineral topsoil, and subsoil samples originating from 46 soil profiles spanning a wide range of soil and vegetation types were analysed. We used Rock-Eval pyrolysis, a technique that investigates the thermal stability of OM, as an indicator of OM dynamics.

Our results show a clear divergence in predictors of OM thermal stability in the litter, topsoil, and subsoil layers. The composition of OM correlated with its thermal stability in the litter layer but not in mineral soil horizons, where the supply rate of fresh organic material and the physical and chemical characteristics of the pedogenic environment appeared important instead. This study offers direct confirmation that soil OM dynamics are influenced by different ecosystem properties in each soil layer. This has important implications for our understanding of carbon cycling in soils under a changing climate.

## 1. Introduction

Soil organic matter (OM) provides essential ecosystem services as it contributes to soil fertility, water quality and retention, biodiversity, resistance to soil erosion, and could play a fundamental role in the mitigation of climate change (Adhikari and Hartemink, 2016). Therefore, it is necessary to understand the mechanisms governing its stability, namely its preservation from mineralisation (Plante et al., 2011; Sollins et al., 1996; von Lützow et al., 2006) in order to maintain soil OM stocks and their associated functions. It was previously widely held that mineralisation rates of soil OM reflected the kinetics of enzymatic reactions and were consequently largely dependent on the intrinsic molecular composition of plant litter entering the soil system (Davidson and Janssens, 2006). This concept has been formalised under the term “selective degradation” (Sollins et al., 1996), and assumed that soil microorganisms preferentially decomposed the inherently labile components of OM, causing the accrual of recalcitrant components (Aber et al., 1990; Melillo et al., 1982). Recent studies have however questioned the idea that organic molecules could be inherently “stable” or “recalcitrant” (Lehmann and Kleber, 2015; Marschner et al., 2008) by showing that potentially persistent organic molecules, such as lignin, could be mineralised relatively quickly in soils (Gleixner et al., 1999, 2002; Heim and Schmidt, 2007). Contrarily, supposedly labile compounds, such as polysaccharides and proteins, can persist in soil for several decades, centuries or even millennia before being mineralised (Derrien et al., 2006; Gleixner et al., 1999, 2002). These long residence times can be in large part attributed to protection or stabilisation by soil minerals (Gleixner et al., 2002; Spielvogel et al., 2008). These recent findings have led to the proposal of a new paradigm, conceptualised by Schmidt et al. (2011). It suggests that selective degradation only plays an essential role in the initial stages of litter decomposition on the soil surface, while its importance becomes marginal when organic material

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is incorporated into the mineral soil. In the mineral soil, OM decomposition rates would instead mainly be driven by its spatial accessibility to microorganisms, their enzymes and the necessary compounds of decomposition (mainly oxygen and moisture), and by the type and number of interactions established with mineral surfaces (Lehmann and Kleber, 2015; Schimel and Schaeffer, 2012; Sollins et al., 1996; von Lützow et al., 2006). OM stability in the mineral soil would thus be mainly governed by ecosystem properties such as climate, soil texture, mineralogy and geochemistry (see synthesis by Schmidt et al., 2011 and references therein).

Even though considerably high proportions (between 30 and 63%) of carbon (C) are stored in the subsoil, between 30 and 100 cm deep (Batjes, 1996), most of the studies on soil OM stabilisation mechanisms have focused on the topsoil (Rumpel and Kögel-Knabner, 2011). This may have resulted in a significant bias in our understanding of drivers of OM stability. Indeed, manipulative laboratory experiments suggest that factors controlling C dynamics in topsoil and subsoil may be substantially different. Fierer et al. (2003) and Salomé et al. (2010) incubated topsoil and subsoil material and found that water potential and supply of fresh organic material were important for surface horizons, while nutrient input, temperature, and the physical accessibility of organic substrates appeared as the main regulatory mechanisms of C mineralisation in the subsurface soil layers. Whether this divergence of controls on soil OM stability is operative under field conditions remains however difficult to evaluate.

Indeed, there is no universally recognised mean to assess soil OM stability. Many different fractionation techniques have been devised based on physical, chemical, or biological properties of OM (see Kögel-Knabner et al., 2008 for a review). Physical and chemical fractionation techniques separate soil OM into operationally-defined pools whose relevance to field-scale OM dynamics remains difficult to assess (Diochon et al., 2016). Investigations that consider the bulk sample without pre-treatment may allow for a more integrative assessment of OM stability. In this respect, biological mineralisation during long-term incubation experiments is generally favoured (Plante et al., 2011), but the inherent soil disturbance and the long durations of incubation required to be fully informative (up to several decades) represent an impediment. Thermal decomposition techniques offer a promising alternative to study soil OM stability. Results from thermal decomposition studies are consistent with those of incubation experiments (Plante et al., 2011) and some physical fractionation schemes (Gregorich et al., 2015; Saenger et al., 2015). The pertinence of thermal techniques is based on the assumption, validated by Plante et al. (2011), that the thermal stability of OM is related to its chemical properties or biological stability, as the activation energy required for thermal bond cleavage correlates to the chemical energy required for enzymatic cleavage (Kögel-Knabner et al., 2008). Schiedung et al. (2017) recently showed that thermal oxidation between 200 and 400 °C was a poor predictor of old (17 years or older) versus recent vegetation inputs. Pyrolysis techniques appear better suited to assess biological stability, as persistent OM tends to disintegrate at higher temperatures than labile OM (Barré et al., 2016). The Rock-Eval pyrolysis technique is now widely employed for routine analysis of OM in soil samples (see Sebag et al., 2016 for a review). This method quantifies total organic and inorganic C contents of a sample (either soil or litter) and provides a wide range of parameters that can be used to evaluate OM composition and its thermal stability. When compared to other methods used to quantify pools of recent or labile C (as assessed using $^{14}$C dating, incubation and physical fractionation), Rock-Eval analysis performed most effectively (Soucémarianadin et al., 2018a; Vinduskova et al., 2015).

In this study, the thermal stability of OM, taken as an indicator of OM dynamics, was measured using Rock-Eval pyrolysis in litter (Oi horizon), topsoil mineral (A horizons), and subsoil mineral layers (including E, B, and C horizons) of 46 subalpine-alpine soil profiles. These soil profiles spanned eight types of vegetation communities and a wide range of soil pH and moisture conditions. The specific aims of this research were (1) to identify the major predictors of OM thermal stability and (2) to assess how their relative importance varied with soil depth. We hypothesised that there would be a clear shift in determinants of OM thermal stability between soil layers, with the influence of plant inputs being restricted to organic layers while the properties of the mineral phase would become prominent at depth.

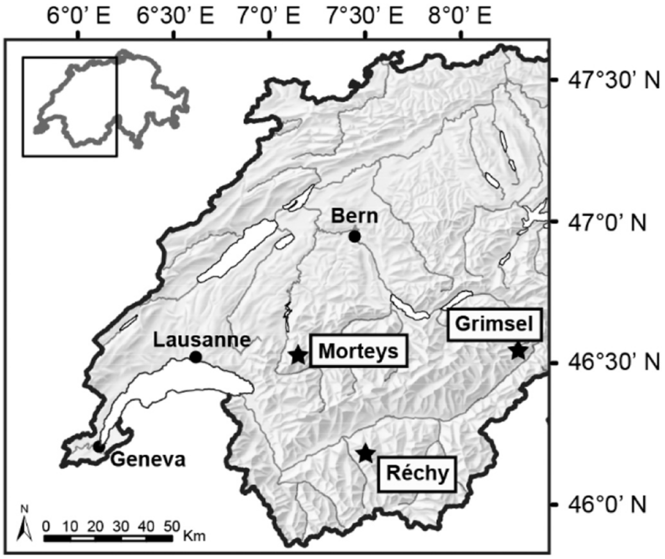

**Fig. 1.** Location of the three study sites (represented by a star) in Switzerland.

## 2. Methods

### 2.1. Sampling design

The 46 soil profiles were selected across three sites of northern and western central Alps in Switzerland (Fig. 1) differing in terms of lithology (Table 1). The Morteys site is underlain by compact limestone and calcareous surficial deposits, while the soil parent materials of the Grimsel site are mainly granite, gneiss, and granodiorite (Oberhänsli et al., 1988). The Réchy area is underlain by a variety of bedrock types including gneiss, mica schist, quartzite, calcschist, marble, and “cornieule” (a dolomite-gypsum greywacke). The three study sites were

**Table 1**
Characteristics of the study sites: coordinates, mean annual temperatures (MAT), mean annual precipitations (MAP), elevation ranges (with median between brackets), vegetation belt, vegetation types present, lithology, and number of soil profiles excavated at each study site. MAT and MAP are extrapolated according to Zimmermann and Kienast (1999) with a 25 m grid cell size.

| | Morteys | Grimsel | Réchy |
|---|---|---|---|
| Latitude | 46°32′N | 46°32′N | 46°10′N |
| Longitude | 7°09′E | 8°16′E | 7°30′E |
| MAT [°C] | 2.1 | −0.44 | −0.53 |
| MAP [mm] | 1650 | 2071 | 1480 |
| Elevation [m] | 1698–2232 (1884) | 2310–2650 (2329) | 2430–2697 (2573) |
| Vegetation belt | Upper subalpine | Lower alpine | Lower alpine |
| Vegetation types | Calcareous grasslands, subalpine pastures, calcareous snowbeds | Siliceous subalpine and alpine grasslands, typical snowbeds | Siliceous alpine grasslands, typical and wet snowbeds, windy ridges |
| Lithology | Limestone | Granite, gneiss, granodiorite | Gneiss, micaschists, quartzite, calcshists, marble, dolomite |
| No. of soil profiles | 18 | 11 | 17 |