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What is the P value of Siberian soils?

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Abstract

Climate change is particularly strong in Northern Eurasia and substantial ecological changes are expected in this wide region. The reshaping and the migration northward of bioclimatic zones may offer opportunities for agriculture development in western and central Siberia. However, the bioclimatic vegetation models currently employed for projections still do not consider soil fertility whereas it is highly critical for plant growth. In the present study, we surveyed the phosphorus (P) status in the south-west of Siberia where soils are developed on loess parent material. We selected six sites differing by pedoclimate conditions and sampled the soil at different depths down to one meter in aspen (*Populus tremula* L.) forest as well as in grassland areas. The P status was assessed by conventional methods and by isotope dilution kinetics. We found that P concentrations and stocks, as well as their distribution through the soil profile, were rather homogeneous at the studied regional scale, although there were some differences among sites (particularly in organic P). The young age of the soils, together with slow kinetics of soil forming processes, have probably not yet conducted to a sufficiently wide range of soil physico-chemical conditions to observe more diverging P status. The comparison of our dataset to similar vegetation contexts on the global scale revealed that the soils of south-western Siberia, and more generally of Northern Eurasia, has often (very) high levels of total, organic and inorganic P. The amount of plant-available P in topsoils, estimated by the isotopically exchangeable phosphate ions, was not particularly high, but intermediate at the global scale. However, large stocks of plant-available P are stored in subsurface layers which have currently low fine root exploration intensities. These results suggest that the P resource is unlikely to constrain vegetation growth and agriculture development in the present and near future conditions.

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1 Introduction

Occupying about 10 million km^{-2} (6.7 % of the global terrestrial land), Siberia has a paramount weight as processes occurring here will have an impact on the global scale. Ranging from 45 to 75° N of latitude, it covers several bioclimatic zones, from south to north: steppe, forest-steppe, sub-taiga, southern taiga, middle taiga, northern taiga, forest tundra and tundra. As the global climate change signal is particularly strong in Northern Eurasia (IPCC, 2013), substantial reshaping of ecosystems is ongoing in the region. The expected increase in average air temperatures will be responsible for longer vegetation growing seasons and frost-free periods, for the melt of permafrost in northern areas and for the modification of soil freeze-thaw cycles in southern areas (Groisman et al., 2012). The intensity and distribution of precipitations may change, leading to differences in fire and hydrological regimes (Shiklomanov and Lammers, 2009; Shkolnik et al., 2010; Soja et al., 2007). These altered physical conditions are expected to modify the composition of the plant communities and the bioclimatic zones of Siberia have been predicted to shift northward and their relative size to change (Jiang et al., 2012; Shuman et al., 2015; Soja et al., 2007; Tchebakova et al., 2009, 2010). In particular, the area occupied by steppe and forest-steppe would increase at the expense of taiga zones. These modifications of ecosystem features may result in alternative land uses (Bergen et al., 2012; Kicklighter et al., 2014). Notably, under future climate conditions, cropping of new species will be possible or existing species may be used on vaster zones than currently in the southern parts of Western and Central Siberia (Kicklighter et al., 2014; Tchebakova et al., 2011). Primary productivity may be enhanced through a "fertilization" effect induced by higher CO_2 levels in the air (Mooney et al., 1991; Norby et al., 2005; Schimel, 1995) along with longer periods sustaining plant growth. However, such projections lack the consideration of other important drivers of plant productivity such as the availability of resources like nutrients and water in soils (Fernández-Martínez et al., 2014; He and Dijkstra, 2014; Oren et al., 2001; Reich et al., 2006a, b, 2014; van Groenigen et al., 2006). Even though there is

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an increasing interest in the study of Siberian ecosystems, functional ecological data remain sparse in the international literature (Gordov and Vaganov, 2010; Groisman and Soja, 2009).

In the domain of biogeochemistry, knowledge about the status of the major nutrients is lacking and potentially misleading assumptions can be made for Siberia. With nitrogen (N), phosphorus (P) is frequently a limiting resource for primary production at the ecosystem scale (Elser et al., 2007; Harpole et al., 2011). It is often considered that P is not the main limiting factor in northern ecosystems (Hedin, 2004; Reich and Oleksyn, 2004). Also, P fertilization in agriculture is barely, if not at all, practised in Siberia. However, a reconsideration of such a paradigm might be necessary in the context of global change. In fact, the increase of atmospheric CO_2 concentration and N deposition, which are in general stronger and faster than any P input, are modifying the CNP stoichiometry of ecosystems (Peñuelas et al., 2013). As a consequence, a progressive shift from N limitation to P limitation or N-P co-limitation can occur (Ågren et al., 2012; Peñuelas et al., 2012; Vitousek et al., 2010). These modifications of biogeochemical cycling at global and regional scales will participate in the driving of ecosystem reshaping, for example through the modification of plant communities adapting to new stoichiometric constraints (Güsewell, 2004). It may also have consequences for agricultural potential at these scales.

In addition, the global resources of P that are used for mineral fertilizer production are limited (Cooper et al., 2011; Cordell et al., 2009). Therefore, enhancing our understanding of P cycles and managing them appropriately at the global scale is highly relevant (Cordell et al., 2011; MacDonald et al., 2011) since the modern terrestrial P cycling is dominated by human activities (Filippelli, 2008). One solution helping to restrict the use of primary P resources would be the development of cropping systems in areas where the soils contains sufficiently plant-available P to prevent the (massive) use of P fertilizers. In this perspective, parts of Siberia are expected to become climatically more suitable for agriculture. Assessing the P status of these Siberian regions is thus of relevance, and this was the main goal of our study.

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In the present study, we proposed to characterize the P status of the soils of SW Siberia, a region characterized by different types of soil along a North–South climatic gradient. We addressed the following questions: (1) how is the P stock structured, in terms of pools and with depth, in the soils of SW Siberia? (2) Which environmental factors control this P status? (3) How can we qualify this P status in comparison to a panel of contrasting pedo-climatic conditions at the global scale? To do so, we selected six sites with contrasting pedo-climatic conditions in SW Siberia and presenting two characteristic vegetation covers, aspen (*Populus tremula*) forests and grasslands. We quantified total P, organic P, phosphate ions in solution and diffusive phosphate ions as a function of time in the soils from these sites. Classical soil analysis methods were combined with an isotopic dilution kinetics method. The size of P pools assessed at Siberian sites were compared with a global dataset compiled from 236 references.

2 Materials and methods

2.1 Site description

Many soils of south-western (SW) Siberia have developed on loess deposits – the Eurasian loess belt covers a broad latitudinal zone between 40 and 60° N – and present favourable texture and mineralogy for plant growth (Chlachula, 2003; Muhs, 2007). Soil formation depends on climate conditions, vegetation cover and can further be shaped by anthropic actions. All of these – climatic conditions, vegetation cover and human activities – differ in intensity essentially along a gradient from south to north. Consequently, from the common origin as loess deposits, the soils in SW Siberia have undergone different development and are classified in the main groupings Chernozems, Phaeozems, and Luvisols. Water-table movements, leaching of carbonates and humus accumulation or organic matter distribution throughout the profile are the most striking features in these soils. The soil forming processes related with soil moisture levels and dynamics as well as the stability of organic matter (mineralization rates in relation with

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temperature and moisture regimes) and the type of vegetation are known to influence the soil P status (Giesler et al., 2002; Miller et al., 2001; Sundqvist et al., 2014; Vincent et al., 2014). We selected six sites in SW Siberia on a transition including forest-steppe and sub-taiga zones. The main site characteristics are given in Table 1 (see also Tables S1–S3 in the Supplement and Brédoire et al., 2015).

All the studied soil profiles developed on a loess parent material and vegetation cover had comparable features in terms of dominant species composition, stand age and low human impact (i.e. no active management for the last decades; Tables S2 and S3 in the Supplement). The main characteristics of the initial loess material are the predominance of coarse-silt particles and clay and the presence of $CaCO_3$, the latter having had different destinies in relation with different soil development processes. In Barnaul (BAR), Chebula (CHE), Krasnozerskoye (KRA) and Salair East (SAE), the main soil forming processes are the formation and accumulation of humus and the leaching of carbonates from it, soils belong to the groups of Chernozems and Phaeozems. In Salair West (SAW) and Tomsk (TOM) soils experience movements of the watertable, with periodical saturation. As a consequence the leaching of carbonates and the washing of clays are important soil forming processes. In addition, the litter is decomposing faster than in forest-steppe and the accumulation of humus is very low. In these two sites, soils belong to the group of Luvisols.

Five of the sites had almost pure aspen (*Populus tremula* L.; Table S2) forest stands along with nearby grassland areas. One site (SAW) only had forest cover with aspen. So there were six sites with forest and five with grassland in our dataset for SW Siberia. All aspen stands had closed canopy.

2.2 Sampling and preparation of the samples

For each vegetation cover on each site, we delimited three study plots (about 300 m^2) spaced by 200–2000 m. Those three plots were considered as replicates. One sampling campaign permitted to sample all sites within three weeks in July 2013. At each plot, we dug a soil pit (on a surface of about 2 by 1 m) down to 120 cm, except in SAE

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grasslands where we reached a dense schist material around 80 cm which prevented us from going deeper than 100 cm. In each soil pit, about 1 kg of soil was sampled horizontally at the depths of 5, 15, 30, 60 and 100 cm ±5 cm with hand tools. Another sample was taken with a cylinder (97 cm^{-3}) to assess soil density.

Litter was sampled over a surface of 30 cm by 40 cm in the vicinity of each soil pit. In this study we defined as "litter" all the dead plant material deposited on the soil surface (senesced leaf litterfall, small branches and senesced understorey vegetation in forests; senesced herbaceous vegetation in grasslands). As a consequence, the material collected in July 2013 resulted mostly from the dead material of the previous vegetation season (2012) and the residues of older seasons, that is to say mainly OL and OF horizons, and eventually OH (at BAR, CHE, KRA and SAE).

Bulk soil samples were air dried until constant weight. After drying, soil samples issued from the same site and the same vegetation cover (i.e. 3 samples per site and per vegetation cover) were pooled and sieved at 2 mm to remove stones and coarse roots. Such soil preparation was reported to affect biogeochemical processes only at a low magnitude (Černohlávková et al., 2009; Chapman et al., 1997). Soil density samples were not pooled. They were oven dried at 105 °C for 48 h and stones were removed when present (i.e. only in deep horizons of SAE).

Bulk litter samples were oven dried at 60 °C until constant weight. They were then pooled by site and by vegetation cover and the composite samples (3 per site and per vegetation cover) were ground before chemical analyses except for the isotopic dilution.

2.3 Physico-chemical analyses

2.3.1 Main soil properties

The French standard methods (Association Française de NORmalisation; AFNOR, 1999) were used for most of the physico-chemical soil analyses. For soil texture, the five-size fractions for clay (< 2 µm diameter), fine loam (2–20 µm), coarse loam (20–

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50 µm), fine sand (50–200 µm), and coarse sand (200–2000 µm) were assessed after decarbonation (NF X 31–107). Soil pH–H_2O was determined in a water/soil suspension with a mass-to volume ratio of 1 g: 2.5 mL (NF ISO 10390). Total organic C and N contents were determined by dry combustion with oxygen (NF ISO 10694 and NF ISO 13878, respectively). Total calcium carbonate contents were assessed with a volumetric method (NF X 31–105). Poorly crystalline aluminium (Al) and iron (Fe) oxides were extracted with an ammonium oxalate solution (McKeague and Day, 1966).

2.3.2 Total, organic and inorganic P

Total P concentrations (P_{tot}, in µg g^{-1} soil) were determined, after grinding, by ICP following wet digestion with concentrated fluoric (HF) and perchloric acids after calcination at 450 °C based on a normalized procedure (AFNOR NF X 31–147; AFNOR, 1999). Total soil organic P concentrations (P_{org}, in µg g^{-1} soil) were determined as the difference of P extracted with H_2SO_4 in ignited and non-ignited soil samples (2 g of dry soil for 70 mL of 0.2 N H_2SO_4; Saunders and Williams, 1955), concentrations were determined with a green malachite colorimetric method (van Veldhoven and Mannaerts, 1987). Total inorganic P concentrations (P_{inorg}, in µg g^{-1} soil) were subsequently calculated as the difference between P_{tot} and P_{org}.

2.3.3 Plant-available phosphate ions

Plants take up P as ions from the soil solution. Thus, a good way of estimating a realistic plant-available P pool in the soil is to quantify both the concentration of phosphate ions in solution and the capacity of the solid phase to maintain this concentration.

To do so, we quantified the phosphate ions in the soil solution (*Cp* in µg mL^{-1} soil solution or *Qw* in µg g^{-1} soil) and the diffusive phosphate ions at the solid–solution interface (*Pr* in µg g^{-1} soil). *Pr* is the quantity of phosphate ions that can be exchanged between solid constituents (ions are adsorbed on soil particles) under a gradient of concentration. *Pr* results from molecular agitation, it can be considered somehow as a "P

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buffering capacity". The sum of *Pr* and *Qw*, that is the isotopically exchangeable phosphate ions (*E*), is considered as a good proxy for the gross amount of plant-available P (Fardeau, 1996; Morel and Plenchette, 1994). *Cp* and *Pr* were determined by an isotopic dilution kinetics method (Fardeau, 1996; Frossard and Sinaj, 1997; Frossard et al., 2011) as described below.

For each litter or mineral soil sample, five suspensions (1 g of litter or soil with 10 mL of deionized water) were equilibrated for 16 h on a roller (40 cycles min^{-1}) at 20 °C (this temperature is commonly reported in the literature and is close to the average temperature of the soil at 20 cm in our study sites in summer, Table S1). Toluene (0.1 mL; M. Lineres, unpublished results) was added at the beginning in the suspension in order to stop microbial activity. This biocide does not affect P biochemical processes (Bünemann et al., 2007). The phosphate ions in solution of the pre-equilibrated suspensions were labelled introducing carrier-free ^{32}P ions in a negligible concentration but knowing the amount of radioactivity introduced (*R*). Suspensions were then sampled with a plastic syringe after 4, 10, 40, 100 and 400 min and filtered on a membrane at 0.2 µm. Then, we quantified both *Cp* and radioactivity remaining in the filtered solution at the time of sampling ($r(t)$). *Cp* was determined with a green malachite colorimetric method (van Veldhoven and Mannaerts, 1987) and *Qw* calculated using the volume of water (*V* in mL) and the mass of litter or soil (m_s in g):

$$Qw = Cp \times \frac{V}{m_s} \quad (1)$$

For each sample, *Cp* was not impacted by the sampling time of the isotopic dilution method (Fig. S1 in the Supplement). The radioactivity remaining in the filtered solution ($r(t)$) was determined in a counter (Packard TR 1100) using a liquid scintillation cocktail. In the steady state conditions of the suspension (*Cp* constant), the gross transfer of phosphate ions from the solid constituents to the solution is equal to the gross transfer of phosphate ions from the solution onto the solid constituents. We assumed that no isotopic discrimination occurs between the two P isotopes (^{31}P ions and ^{32}P ions) during the transfers between the solution and the solid phases. The amount of unlabeled

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phosphate ions newly transferred from the solid constituents to the solution ($Pr(t)$) was then calculated from Qw and $r(t)$ values following the principle of isotopic dilution (R is diluted in E).

$$\frac{R}{E} = \frac{r(t)}{Qw} = \frac{R - r(t)}{Pr(t)} \tag{2}$$

Rearranging Eq. (2) gives:

$$Pr(t) = \frac{Qw \times (R - r(t))}{r(t)} = Qw\left(\frac{1}{r(t)/R} - 1\right) \tag{3}$$

where $\frac{r(t)}{R}$ (dimensionless) is the isotopic dilution ratio.

The theoretical Eq. (4), adapted from Fardeau (1993, 1996) was used to closely fit the experimental values of $\frac{r(t)}{R}$ as a function of isotopic dilution time:

$$\frac{r(t)}{R} = m\left(t + m^{\frac{1}{n}}\right)^{-n} \quad \text{for} \quad \frac{r(t)}{R} \geq \frac{r(\infty)}{R} \tag{4}$$

where m and n are fitting parameters and $\frac{r(\infty)}{R}$ corresponds to the maximum possible dilution of the isotope considering that all inorganic P can take part in the isotopic dilution. The value of $\frac{r(\infty)}{R}$ tends toward $\frac{Qw}{P_{inorg}}$ (Fardeau, 1993; Frossard et al., 2011). The parameter m, which is the fraction of radioactivity remaining in solution after 1 min ($\frac{r(1\,\text{min})}{R}$), accounts for the immediate physico-chemical reactions while the parameter n accounts for the slow ones (Fardeau et al., 1991; Fardeau, 1993). The quality of the fit for Eq. (4) is shown in Fig. S2 in the Supplement and the values of m and n are provided in Table S5 in the Supplement.

Combining Eqs. (3) and (4), we can derive the value of Pr over time, each value corresponding to a pool of P more or less rapidly available to plants. The number and the size of such pools can be defined considering plant functioning (Fardeau, 1993).

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In this study, we computed the values of *Pr* for 1 day, 1 week, 1 month and 3 months. While 1 day is the mean duration for active root uptake, 3 months is approximately the duration of the vegetation season in south-western Siberia and we might expect this to fit with intense root activity.

2.4 Data handling and statistics

Five soil layers were defined between 0 and 120 cm according to soil horizon description in each soil pit (horizons were merged or divided in order to have 5 layers corresponding to the 5 sample depths, the studied profiles presented between 4 and 7 horizons, the mean number of horizons of the 33 studied profiles is 5). Assuming elemental concentrations and soil densities measured in each of the five defined horizons were representative of the entire horizon, we computed stock (in $Mg\,ha^{-1}$) of each P pool using mean soil densities and horizon thicknesses:

$$\text{stock} = \frac{1}{10\,000} \times [P] \times d \times h \quad (5)$$

where $[P]$ is the concentration of the P pool (in $\mu g\,g^{-1}$), d the soil density (in $g\,cm^{-3}$) and h the thickness of the soil horizon in cm. In litter, the P stocks were computed (in $Mg\,ha^{-1}$) using the mass and the surface sampled:

$$\text{stock} = \frac{1}{10\,000} \times [P] \times m_{\text{litter}} \quad (6)$$

where $[P]$ is the concentration of the P pool (in $\mu g\,g^{-1}$) and m_{litter} the mass of litter (in $g\,m^{-2}$). We used the limit of −20 cm to distinguish between top- and subsoil to fit with most of agronomic studies but also with the zone where most of fine root exploration occurs in SW Siberia (Brédoire et al., 2015).

In order to get an idea of the structure of the P stocks in the studied soils, we computed the relative proportions of the different measured P pools as a fraction of P_{tot}.

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P_{tot} is the sum of P_{org} and P_{inorg}. Since a biocide was added in the suspension, mineralization was stopped and we only measured physico-chemical processes. Thus, all exchangeable P (the sum of *Qw* and *Pr*) is part of P_{inorg}. *Pr* being calculated as a function of time and the maximum time considered in this study being 3 months, the potentially remaining fraction of P_{inorg} is considered as non-diffusive, or diffusive in more than 3 months. We note that when considering two values of *Pr* computed at different times, the pool of exchangeable P computed at the longer time includes the one computed at a shorter time.

Relations between P parameters and soil physico-chemical properties were investigated through the computation of Spearman's rank correlation coefficients, scatter plots and (non-)linear regressions. Soil physico-chemical properties varied with soil depth (Table S4 in the Supplement) as well as the investigated P parameters (Table 2). Thus, we looked for correlations in each soil layer separately in order to avoid covariation and interdependencies issues (Table S6 in the Supplement). We also analyzed correlations with fine root (diameter < 0.8 mm) length density (FRLD) and fine root mass density (FRMD) measured in the same soil pits and at the same soil depths as for the soil physico-chemical properties (Brédoire et al., 2015).

Since analyses were made on composite samples we did not quantify the variability of our measurements at the level of the site for given vegetation cover and soil depth. Thus, differences between sites were not tested through formal statistical tests. However, we calculated the coefficient of variation (ratio of the standard deviation to the mean) for each layer and vegetation cover.

All data management, (non-)linear regressions and statistical analyses (correlation coefficients and their significance), were performed with R 3.2.1 (R Core Team, 2015).

2.5 Comparison at the global scale

To compare the phosphorus status of our study sites with other grassland or forest ecosystems and with croplands, we made a compilation of data on different P fractions in soils. In practice, we used different requests involving keywords such as "soil",

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"phosph*", "total content", "isotopic dilution", "isotopically exchangeable P", "grassland", "forest", "woodland", etc. These requests were carried out both in Web of Science and Google Scholar. To derive the pools of diffusive and isotopically exchangeable phosphate ions, we selected all publications using the same isotopic dilution procedures as in the present study (i.e. Fardeau's procedure; Fardeau, 1993, 1996). Additional publications were selected in order to improve the geographical coverage for total and organic P. In particular, we examined all the tables of contents of the *Soviet Soil Science* and the *Eurasian Soil Science* journals to better cover Northern Eurasian ecosystems. Based on all the selected references, we compiled a dataset of different P fractions (total P, organic P, phosphate ions in soil solution, diffusive phosphate ions and isotopically exchangeable phosphate ions) in soils of grasslands, forests, or croplands. This database contained P values for up to 373 distinct sites depending on the P fraction, the geographical scale and the vegetation type studied, and collected from 236 references. This database was representative of the soils of the world as shown by the geographical distribution of compiled sites (Fig. 1), even though the studies using the isotopic dilution kinetics method in forest were sparse. In particular, out of the 116 forest study sites present in the compilation – with values for inorganic P, organic P, phosphate ions in solution, diffusive phosphate ions and isotopically exchangeable phosphate ions –, 106 are located in France. Nevertheless, France has very diverse soil and geology contexts. The most represented soil types are Podzols, Cambisols and Luvisols, but Planosols, Leptosols, Calcisols, Arenosols, Regosols and Andosols are also present in some sites (IUSS Working Group WRB, 2014). The main parent materials are calcareous formations, eruptive and metamorphic rocks, sandstone, detritic and weathered formations. Among these 106 French sites, 50 are hardwood forests (2 species) and 56 are conifer forests (5 species). Therefore, our dataset was representative of forests at the global scale for soil total P, and representative of very diverse temperate forests for isotopic P data. The references of the data compilation are provided in the Supplement.

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3 Results

3.1 Quantification of P pools

The concentrations of total P (P_{tot}; 694–1095 µg g^{-1} at −5 cm, 319–694 µg g^{-1} at −100 cm), organic P (P_{org}; 389–774 µg g^{-1} at −5 cm, 37–79 µg g^{-1} at −100 cm) and phosphate ions in solution (*Qw*; 2–22 µg g^{-1} at −5 cm, 0.1–0.4 µg g^{-1} at −100 cm) decreased with depth over the 1 m profile in all the studied sites for both forest and grassland (Table 2). The litter layer presented the highest concentrations for these pools; *Qw* being one to two order(s) of magnitude more concentrated in the litter (223–638 µg g^{-1}) than in the upper soil layer (2–22 µg g^{-1} at −5 cm). No systematic variation with depth was found over the profile for inorganic P (P_{inorg}; 296–626 µg g^{-1} at −5 cm, 282–616 µg g^{-1} at −100 cm) and diffusive phosphate ions (e.g. *Pr*(1 day); 16–56 µg g^{-1} at −5 cm, 2–67 µg g^{-1} at −100 cm; Table 2). However, *Pr*(1 day) was decreasing over the three first mineral soil layers except for the grassland in SAE.

We computed stocks (Mg ha^{-1}) for the different P pools (Table 3). With the exception of *Qw*, the subsoil contributed the most to the total stocks (72–85 % of P_{tot}, 64–73 % of P_{org} excluding SAE, 82–90 % of P_{inorg} and 67–94 % of *Pr*(1 day)). The three layers considered (litter, topsoil and subsoil) contributed almost equally to the total stock of *Qw* (respectively, 10–56, 26–65 and 7–49 %). All sites presented values in the same order of magnitude for a given P pool and a given layer. The values for forest and grassland were also close. One notable difference occurred at the site SAE, where soil P pools were lower in forest than in grassland and where the pools in forest were lower than in the other sites. KRA presented the highest stocks in litter for all P pools, however, its stocks in the topsoil were the lowest (except for *Qw* and *Pr*(1 day) in grassland) and they were also relatively low in the subsoil.

For each layer, we calculated the relative contribution of each P pool to total P (Fig. 2). With the exception of one grassland site (TOM), P_{org} accounted for more than 50 % of P_{tot} in the litter layer. The concentration of phosphate ions in solution (*Qw*) rep-

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resented 20–38 % of P_{tot} in the litter layer with close values for forest and grassland at each site. One site (TOM) presented much higher values of *Qw* in litter, reaching 45 and 67 % of P_{tot} in forest and grassland respectively. All the studied sites, and whatever the vegetation cover, exhibited the same pattern along the mineral soil profile. The relative proportion of P_{org} decreased while the relative proportion of P_{inorg} increased with depth. In the two upper soil layers the distribution of P pools was very comparable. P_{org} accounted for 34–71 % of P_{tot} and *Pr* (3 months) for 3–13 %. *Qw* represented up to 2 % of P_{tot} at −5 cm, dropped below 0.1 % at −30 cm and was around 0.01 % at −100 cm. The proportions of *Pr* tended to be higher in forest than in grassland. Below −15 cm, the proportion of P_{org} decreased down to 6–19 % at −100 cm. In the deepest layers, the proportions of *Pr* tended to be higher than in the first two ones with notable exceptions: extremely low values at −100 cm in SAE, and *Pr* (3 months) representing 100 % of P_{inorg} at −60 cm in TOM grassland.

3.2 Relations between P pools and environmental parameters

We tested the correlations between P parameters and the main soil physico-chemical properties (Table S6 and Fig. S3 in the Supplement). P_{inorg} was significantly correlated with P_{tot} (Spearman's rank correlation coefficients ranging between 0.627 and 0.989). This was also observed between *n* and pH at all depths except at −15 and −100 cm (−0.636 to −0.793). In the three top layers, P_{org} was significantly correlated with organic C (0.682 to 0.843) and *m* with *Qw* (0.609 to 0.855). In the two deepest layers, *Pr* and isotopically exchangeable phosphate ions (*E*) were significantly correlated with the clay fraction (0.782 and 0.852). They were also negatively correlated with $CaCO_3$ (−0.649) at −60 cm (Table S6) but this was driven by one point very depleted in $CaCO_3$ and with very high *Pr* and *E* (Fig. S3). A few correlations were found with Al and Fe oxides: with P_{org} at −30 cm (−0.636) and with *n* at −60 cm (0.718).

A few relationships between fine root densities and P pools were significant (Table S6 and Fig. S3). At −15 cm, fine root length density (FRLD) was significantly and negatively correlated with *Qw* (−0.636), *m* (−0.764), *Pr* (1 day) (−0.691) and *E* (1 day)

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(−0.736). At −30 cm, FRLD was significantly and negatively correlated with *n* (−0.773) and fine root mass density (FRMD) with *n* (−0.655), *Pr* (0.636) and *E* (0.618).

No relation was found between the different variables of the P status and any of the climate parameters presented in Table S1 (data not shown).

3.3 Comparison at the global scale

Total P concentrations in topsoil (first 20 cm of the soil) ranged at the global scale between 62 and 2480 $\mu g\,g^{-1}$ in croplands, between 19 and 3090 $\mu g\,g^{-1}$ in forests and between 32 and 3548 $\mu g\,g^{-1}$ in grasslands (Fig. 3). Our measurements in SW Siberia ranged between 345 and 770 $\mu g\,g^{-1}$ in forests and between 481 and 741 $\mu g\,g^{-1}$ in grasslands, these values were close and above the global upper quartile for forests and between the global median and the upper quartile for grasslands. Compared with global cropland values these Siberian concentrations were mostly above the upper quartile. Restricting the domain of comparison to Northern Eurasia, SW Siberian soils ranged between the lower and the upper quartiles for forests and between the median and the upper quartile for grasslands, indicating that our sites are representative of Northern Eurasia. In the subsoil (−20 to −100 cm), less points were available at the global and at the Northern Eurasian scales for comparison. However, our SW Siberian forest and grassland soils occupied wider ranges in comparison to the corresponding vegetation types: from below the median to above the upper quartile at the global scale and a similar range of values to Northern Eurasia.

The concentrations of P_{org}, P_{inorg} and phosphate ions in solution (*Cp*) in the topsoil of the studied SW Siberian forests were generally above the upper quartile – comparison essentially with French forests of contrasting species, soil and geology contexts – (Fig. 4). In grassland, the SW Siberian values were mostly comprised between the global median and upper quartile. Compared with global cropland ranges, our measurements were around and above the upper quartile for P_{tot} and P_{org}, mainly comprised below the upper quartile for P_{inorg} and from the median to above the upper quartile for *Cp*. Interestingly, the proportion of P_{org} (% of P_{tot}) measured varied quite a lot on the

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range reported on the comparison scales, particularly in forest where SW Siberian values varied from below the lower quartile to above the upper quartile. By contrast to the other P pools, the concentrations of *Pr* (1 day) and *E* (1 day) were more moderate: they ranged between the median and the upper quartile in forest and between the lower quartile and the median in grasslands, these values being lower than the global cropland median.

4 Discussion

4.1 A relatively homogeneous P status

All the sites, for both aspen forest and grassland, presented a similar distribution of the P pools through the soil profile. Total P concentrations decreased with depth, mainly in relation with the decrease in the concentration of organic P (Table 2). The stocks varied in the same order of magnitude between sites and, among sites, between contrasting vegetation cover types. The concentrations and the stocks we computed were close to the ones reported by Achat et al. (2013a). These authors reported standard errors of 1–27 % for the concentrations of total P, organic P, inorganic P, phosphate ions in soil solution and diffusive phosphate ions, with three replicates per condition in two sites of the same region. Assuming a similar spatial variability, the concentrations and stocks we measured or computed for given soil layers appeared relatively homogeneous (values in the same order of magnitude without notable outliers) at the regional scale.

Despite lower concentrations of the P pools in the subsoil, it contributed the most to the total stocks computed over 1 m (Table 3) because of its higher thickness. For the same reason, at the scale of the soil profile, inorganic P represented far more than 50 % of the stock of total P (Table 3) while organic P concentrations represented a high proportion of total P in the litter and in the three first soil layers (Table 2). Over the profile, the P stock in the studied soils can be qualified as predominantly mineral over 1 m.

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4.2 Environmental factors controlling the regional P status

In spite of small variations in the current P status in the investigated soils, we found this status to be impacted by a set of variables. At first, we observed that the amount of P is highly dependent on the amount of inorganic P, particularly in the subsurface layers of the studied sites (Fig. 2 and Table 3). Biogeochemical cycling and soil development processes explain the P status over depth. In topsoil, organic P represented a large part of total P (Fig. 2) and was related to organic C (Table S6 and Fig. S3). This is a direct consequence of P uptake, immobilization in plant tissues, followed by litterfall and subsequent accumulation in the top horizons (Barber, 1995). Clay minerals (on their edges), carbonates and organic matter have surfaces presenting positive charges that are reactive with phosphate ions (Gérard, 2016; Hinsinger, 2001; Parfitt, 1978). In topsoil, the preponderance of organic P suggests that microbial processes may play an important role in the plant-availability of P through the release of phosphate ions by mineralization. On the contrary, in deep horizons P plant-availability is principally explained by mineral phases such as the clay fraction (Table S6 and Fig. S3). Contrary to other case studies (Achat et al., 2011; do Carmo Horta and Torrent, 2007; Tran et al., 1988; Walbridge et al., 1991), we found only a few relations with Al and Fe oxides concentrations. These oxides also present positive charges that are known to be reactive with phosphate ions (Achat et al., 2011; Regelink et al., 2015).

The inspection of the correlations involving the parameters *m* and *n* provides further insights on the drivers of phosphate ions exchange at the solid–solution interface. Correlations with *m*, which is the fraction of radioactivity remaining after one minute in the isotopic dilution, informs about rapid processes, while correlations with *n* are considered to be indicative of processes driving slow exchange reactions. Both parameters were related to the phosphate concentration in solution (*Qw*). Generally *m* increased with *Qw* while *n* decreased (Table S6 and Fig. S3), in agreement with previous studies (Achat et al., 2009, 2013a; Fardeau, 1993; Morel et al., 2000). The dynamics of phosphate ions (slow reactions in the present study) were further dependent on pH, which

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modifies the charge of reactive solid surfaces and the speciation of phosphate ions (Barrow, 1983; Hinsinger, 2001; Strauss et al., 1997a, b; Ziadi et al., 2013). In addition, we found slight and secondary effects of Al and Fe oxides on slow phosphate ions exchange reactions: residual values of the parameter *n* increased with the increase in oxides content (data not shown). This fit with a preliminary study in SW Siberia (see more details on the effects of Al and Fe oxides on parameter *n* in Achat et al., 2013a).

Nevertheless, at some sites, a few layers did not exhibit the general features of the P status described above. Soil formation processes and soil physico-chemical properties also explain that we observed such "outliers". Despite overall slight effects only, soil content in Al and Fe oxides had some visible influence on soil P at the scale of some soil profiles. It was the case at TOM, a site with a watertable close to the topsoil. The periodical watertable movements may be responsible for the relatively stronger accumulation of clays and oxides in deeper soil layers (Table S4) and contribute to the higher concentrations and proportions of diffusive phosphate ions of the soil (especially at −60 cm in grassland; Table S4). At KRA, the accumulation of $CaCO_3$ (Table S4) could be responsible for high levels of diffusive phosphate ions in the subsoil. In fact, in alkaline soils such as in KRA, phosphate ions tend to precipitate with Ca cations which have an increasing solubility at pH above 8 (Hinsinger, 2001; Kuo and Lotse, 1972). At SAE, the schist material underlying the loess deposit (below −80 cm) is probably responsible for low P pool concentrations (particularly in forest) and extremely low proportions of diffusive phosphate ions in the deep layers compared to the other sites.

The restricted number of significant correlations – between P pools or isotopic dilution parameters (*m* and *n*) and soil physico-chemical properties – we identified in our study is not necessarily indicative of an absence of control on the P status. It may simply reflect that the values of the tested soil variables and P pools stand within a restricted range (same order of magnitude or difference of only one order; Table 2 and S4). Soils of the SW part of Siberia are indeed relatively homogeneous. They have developed on loess material deposited during the Quaternary era, mainly during the two last glaciation periods (Chlachula, 2003; Muhs, 2007) and despite some contrasted

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climate conditions, they have not been sufficiently impacted by diverging pedogenetic processes. Additionally, soil forming processes are expected to be relatively slow in such a dry and cold region (Jenny, 1941).

4.3 High levels of total P fractions but medium ones for plant-available P

In a general manner, the studied SW Siberian soils presented for total P, organic P, inorganic P and phosphate ions in soil solution very high levels in forest and high levels in grassland when compared with our compilation of data at the global scale – or diverse soil and geological contexts mainly in France, for all P pools but total P, in forests – (Figs. 3 and 4). In addition, the high level of total P stocks might be generalizable to Northern Eurasia, at least for the soils developed in the loess belt. Of course, more field measurements are requested to verify this statement, particularly in the vast zone currently covered by taiga and not in the loess belt. On the other hand, this result, if confirmed, would be of primary importance in a context of global change and of tensions on the resources of P for agriculture.

However, we noted that these relatively high concentrations of total P in SW Siberian soils did not automatically indicate a high P availability for plant nutrition. In fact, an important parameter is the ability of the soil to refill a depleted soil solution (e.g. due to root uptake) with phosphate ions. This "P buffering capacity" is assimilated to the quantity of diffusive phosphate ions between the solid and the liquid phases of soil. Contrarily to the other measured P pools in the studied SW Siberian soils, the concentrations of diffusive phosphate ions in the topsoil were not so high, in comparison to global levels (although not being very low; Fig. 4).

In French forests, the sum of Al and Fe oxides ranges from from 4.5 to 1157.7 $mmol\,kg^{-1}$ and pH from 3.6 to 8.3 (data compilation of 106 sites, unpublished). In comparison, our SW Siberian soils have a low sum of Al and Fe oxides (68.44–184.08 $mmol\,kg^{-1}$) – and the narrow range of values explains why we found only a few correlations between P pools and oxides – and a very high pH (5.37–7.16, Table S4). This very high pH is probably partly responsible for a low reactivity of phosphate ions,

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notably because the number of positive charges decreases with increasing pH (Barrow, 1983; Hinsinger, 2001; Ziadi et al., 2013). Coupled with a low amount of oxides (i.e. fixation sites), this might explain the average values of diffusive (*Pr*) and isotopically exchangeable (*E*) phosphate ions in the studied SW Siberian soils while total pools were (very) high.

Following the conceptual model of (Walker and Syers, 1976), which describes the changes in the forms and amounts of P pools with time, and according to the comments we made in the sections above on the regional homogeneity of the P status, we concluded that these SW Siberian soils are probably in the early stages of soil development. Such a stage is characterized by the build-up of an appreciable organic P stock, but also by a still large stock of primary inorganic P. Thus, it may still exist a high potential of primary mineral weathering in these soils. Moreover, the mineralization of organic matter is another source of phosphate ions for the refilling of the soil P buffering capacity (Achat et al., 2013b; Bünemann, 2015). The study of the kinetics of these mechanisms is relatively difficult and was not done in the scope of this study. However, they could be of importance as they are likely to be impacted by global change. Organic matter mineralization would mainly depend on temperature and moisture (Bengtson et al., 2005; Paul et al., 2002). Mineral alteration would mainly depend on temperature and pH (Augusto et al., 2000; Drever, 1994).

The absence of correlation between fine root densities and P pools (Table S6) suggests that root exploration is not related with P mining. In addition, the relatively low N : P ratios measured in the litter layers (9–14; Table S4) as well as in aspen green leaves (9–12, data not shown) suggest P is unlikely to be the primary limiting nutrient for plant nutrition and litter decomposition, that would be N (Aerts and Chapin, 1999; Reich and Oleksyn, 2004; Güsewell and Gessner, 2009). Thus the availability of P probably not constrain much plant growth in the studied ecosystems. This conclusion is in line with the review made by (Smurygin, 1974) about fertilization experiments in the former USSR.

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Will this P status be sufficient to fulfil future plant requirements in the context of global change? In the speculative situation where topsoils are depleted by intense biomass exports, our results suggest that the large P stocks in deeper soil layers could sustain the demand (Table 3). This would imply a deepening of plant fine root systems. Such a deepening of fine root systems have already been observed with ongoing global change: in relation with the increase in atmospheric CO_2 concentrations (Iversen, 2010) or with the lengthening of vegetation growing seasons (Lempereur et al., 2015; Majdi and Öhrvik, 2004). In SW Siberia, we hypothesize a deepening in fine root systems would more likely be driven by other resources, particularly water in the steppe and forest-steppe zones (Brédoire et al., 2015).

5 Conclusions

This study revealed that the concentrations, the stocks of the different P pools measured and their distribution over the soil profile were relatively homogeneous at the scale of SW Siberia although there were some differences among sites (mainly organic P) maybe due to varying microbial activity and slight differences in physico-chemical soil properties. In this region, we argue that the young age of the soils developed on loess parent material, coupled with slow kinetics of pedogenesis, have probably not yet conducted to a sufficiently wide range of soil physico-chemical conditions to observe more diverging P status. The comparison of these Siberian P levels to similar vegetation contexts on the global scale revealed high to very high levels of total, organic and inorganic P in topsoils. These results seems to be generalizable to Northern Eurasia but additional measurements are requested to verify this statement. The amount of plant-available P in topsoils, evaluated as isotopically exchangeable phosphate ions, was intermediate at the global scale. However, large stocks of isotopically exchangeable phosphate ions are stored in the subsurface layers where fine root exploration is currently low. These results suggest that the P resource is unlikely to constrain vegetation growth and agriculture development in the present and near future conditions.

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Table 1. Main characteristics of the study sites. Additional informations are provided in Tables S1–S3.

Site ID	BAR	CHE	KRA	SAE	SAW	TOM
Geographical characteristics						
Name[a]	Barnaul	Chebula	Krasnozerskoye	Salair East	Salair West	Tomsk
Ecological zone	Forest-steppe (southern part)	Forest-steppe (northern part)	Steppe to Forest-steppe	Sub-taiga to Forest-steppe	"Blackish taiga"	Sub-taiga
Latitude (° N)	53.41	55.55	53.59	54.39	54.18	56.30
Longitude (° E)	83.47	84.00	79.14	85.75	85.17	85.43
Elevation (m a.s.l.)	221	186	141	305	358	232
Climate characteristics (annual mean 1981–2010)						
Air temperature (°C)	2.7	1.3	2.9	2.3	1.2	0.9
Precipitation (mm)	431.5	509.8	324.5	432.3	453.0	566.5
Soil classification						
Soil in forest	Haplic Phaeozem	Haplic Phaeozem	Phaeozem	Leptic Phaeozem	Haplic Luvisol	Albic Luvisol
Soil in grassland	Calcic Chernozem	Haplic Phaeozem	Calcic Hortic Chernozem	Leptic Phaeozem		Albic Luvisol
Forest stand characteristics (mean values)						
Density (tree ha^{-1})	1664	387	767	1883	1144	1139
DBH (cm)	14.9	33.9	26.3	13.7	22.8	21.4
Height (m)	11.2	28.0	18.7	15.7	24.8	18.2
Age (years)	27	62	51	21	47	56

[a] Name of the closest city or name of the local area.

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Table 2. Phosphorus concentrations of different pools measured in litter and soil layers of south-western Siberia. Concentrations are expressed in $\mu g\,P\,g^{-1}$ soil (or litter), depth is in cm. "Litter" means all the dead plant material deposited on the soil surface (senesced leaf litterfall, small branches and senesced understorey vegetation in forests; senesced herbaceous vegetation in grasslands) that is to say mainly OL and OF horizons, and eventually OH (at BAR, CHE, KRA and SAE) at the date of sampling.

P pool	Depth (cm)	Forest							Grassland					
		BAR	CHE	KRA	SAE	SAW	TOM	CV	BAR	CHE	KRA	SAE	TOM	CV
Total P	Litter	1235.6	1174.5	1318.6	1231.3	1515.1	1011.8	**13**	1165.8	1340.4	1122.1	1318.6	953.1	**13**
	−5	1017.3	851.4	729.2	1095.9	1039.2	759.7	**17**	877.6	956.2	847.0	921.3	694.2	**12**
	−15	951.8	663.7	615.6	908.2	864.5	676.8	**19**	864.5	938.7	698.6	873.2	624.4	**17**
	−30	960.6	628.7	541.4	416.1	768.5	558.9	**30**	755.4	899.4	628.7	563.2	528.3	**23**
	−60	764.1	676.8	394.7	323.5	585.1	489.0	**31**	593.8	689.9	515.2	611.3	493.4	**14**
	−100	637.5	659.3	408.2	319.2	646.2	528.3	**27**	593.8	694.2	414.4	593.8	537.0	**18**
Organic P	Litter	817.3	760.1	839.5	896.2	910.8	758.9	**8**	800.5	690.4	640.7	855.5	660.0	**13**
	−5	391.6	405.6	418.5	773.7	470.0	436.1	**30**	388.8	539.2	539.5	521.8	398.7	**16**
	−15	319.6	286.8	335.1	614.6	379.4	375.7	**31**	358.3	472.2	413.7	495.6	333.6	**17**
	−30	332.1	208.7	286.4	202.6	284.3	217.5	**21**	276.9	422.0	343.6	100.3	206.1	**46**
	−60	156.8	133.5	96.9	58.2	114.7	137.8	**30**	108.1	90.1	201.4	123.4	89.7	**38**
	−100	50.3	58.8	79.5	37.1	68.5	63.5	**25**	48.6	78.7	45.7	37.0	64.5	**30**
Inorganic P	Litter	418.3	414.4	479.1	335.1	604.3	253.0	**29**	365.3	650.0	481.4	463.1	293.1	**30**
	−5	625.7	445.8	310.7	322.2	569.2	323.7	**32**	488.8	417.0	307.6	399.4	295.5	**21**
	−15	632.2	376.9	280.5	293.6	485.1	301.1	**35**	506.3	466.5	284.9	377.7	290.8	**26**
	−30	628.5	420.1	255.1	213.5	484.2	341.4	**39**	478.5	477.4	285.2	462.9	322.2	**23**
	−60	607.3	543.3	297.8	265.3	470.4	351.2	**33**	485.7	599.7	313.9	487.9	403.7	**23**
	−100	587.1	600.5	328.8	282.1	577.7	464.8	**29**	545.2	615.5	368.7	556.8	472.6	**19**
Phosphate ions in soil solution	Litter	353.2	397.6	406.1	343.8	500.4	457.1	**15**	393.7	520.7	223.2	271.2	637.8	**42**
	−5	10.3	15.8	11.2	22.1	17.9	1.5	**55**	6.4	2.9	13.7	1.9	1.7	**95**
	−15	5.3	2.3	1.5	6.0	3.3	1.6	**57**	1.0	1.3	1.1	1.0	0.7	**23**
	−30	3.6	0.8	0.7	0.4	0.9	0.2	**114**	0.4	0.8	0.8	0.5	0.1	**56**
	−60	1.1	0.3	0.5	0.1	0.2	0.2	**89**	0.2	0.3	0.7	0.7	0.2	**62**
	−100	0.2	0.2	0.2	0.1	0.4	0.3	**44**	0.2	0.2	0.4	0.1	0.1	**50**
Diffusive phosphate ions in 1 day	Litter	88.8	84.2	72.9	80.0	103.9	37.3	**29**	36.6	23.5	59.9	85.8	3.0	**77**
	−5	42.5	56.1	34.9	53.4	48.1	26.3	**26**	31.6	20.3	36.1	21.1	16.2	**33**
	−15	33.1	25.0	17.1	26.7	35.4	26.2	**24**	21.5	15.8	27.3	22.8	18.7	**20**
	−30	29.2	15.3	26.6	15.6	24.6	11.9	**35**	16.6	13.5	17.7	40.8	13.7	**56**
	−60	39.6	38.4	30.5	14.7	25.4	29.9	**31**	20.0	48.5	20.4	13.4	89.4	**82**
	−100	24.3	40.4	32.3	2.2	66.7	57.4	**62**	17.5	35.3	47.4	2.5	36.3	**64**

CV: coefficient of variation (ratio of the standard deviation to the mean, in %).

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Table 2. Continued.

P pool	Depth (cm)	Forest							Grassland					
		BAR	CHE	KRA	SAE	SAW	TOM	**CV**	BAR	CHE	KRA	SAE	TOM	**CV**
Diffusive phosphate ions in 1 week	Litter	88.8	84.2	72.9	80.0	103.9	37.3	**29**	36.6	23.7	74.3	109.8	3.0	**86**
	−5	64.7	92.7	52.0	79.8	74.5	57.3	**22**	47.8	33.1	53.8	33.7	29.4	**27**
	−15	52.9	45.9	29.8	43.5	70.0	51.0	**27**	34.2	26.8	51.3	42.7	39.3	**24**
	−30	46.4	28.8	57.2	29.4	51.5	26.6	**33**	28.4	24.3	30.9	71.0	31.7	**51**
	−60	70.1	73.9	51.1	31.9	53.9	64.4	**27**	35.2	95.9	37.1	20.2	178.7	**89**
	−100	47.0	73.1	61.7	4.0	137.6	91.8	**65**	30.4	65.1	94.3	4.6	64.5	**67**
Diffusive phosphate ions in 3 months	Litter	88.8	84.2	72.9	80.0	103.9	37.3	**29**	36.6	24.0	94.3	143.7	3.0	**96**
	−5	108.7	170.4	84.5	129.2	125.8	155.5	**24**	80.1	61.0	87.1	61.6	62.4	**17**
	−15	95.3	99.1	60.4	79.1	166.4	119.5	**36**	62.4	52.4	116.4	96.2	102.6	**32**
	−30	83.4	65.0	154.6	66.7	133.8	75.6	**39**	57.1	51.5	63.5	146.2	94.5	**48**
	−60	147.4	174.1	100.3	87.9	144.2	175.4	**27**	73.9	234.2	80.4	34.1	403.4	**93**
	−100	111.5	159.3	144.1	8.5	355.3	169.6	**71**	62.4	145.0	231.7	10.2	136.9	**72**

CV: coefficient of variation (ratio of the standard deviation to the mean, in %).

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Table 3. Stocks of the different P pools computed in Mg ha^{-1} for the litter, the topsoil (about 0 to −20 cm), the subsoil (about −20 to −120 cm) and over the profile for the different sites. "Litter" means all the dead plant material deposited on the soil surface (senesced leaf litterfall, small branches and senesced understorey vegetation in forests; senesced herbaceous vegetation in grasslands) that is to say mainly OL and OF horizons, and eventually OH (at BAR, CHE, KRA and SAE) at the date of sampling.

		Forest							Grassland					
P pool	Layer	BAR	CHE	KRA	SAE	SAW	TOM	**CV**	BAR	CHE	KRA	SAE	TOM	**CV**
Total P	Litter	0.037	0.028	0.057	0.024	0.014	0.012	**58**	0.008	0.003	0.026	0.012	0.002	**95**
	Topsoil	2.549	1.749	1.121	1.462	1.729	1.433	**29**	2.338	2.030	1.464	2.089	1.432	**22**
	Subsoil	9.728	8.865	6.247	3.887	10.190	7.287	**31**	9.541	9.682	6.848	7.740	7.366	**16**
	Total	12.313	10.642	7.424	5.374	11.934	8.732	**29**	11.888	11.714	8.338	9.840	8.800	**16**
Organic P	Litter	0.024	0.018	0.036	0.018	0.009	0.009	**55**	0.006	0.001	0.015	0.008	0.001	**89**
	Topsoil	0.894	0.782	0.622	1.003	0.764	0.805	**16**	0.995	1.079	0.896	1.185	0.791	**16**
	Subsoil	2.010	1.520	1.787	0.958	2.075	1.705	**24**	1.809	2.002	1.964	1.043	1.604	**23**
	Total	2.928	2.320	2.445	1.978	2.848	2.518	**14**	2.809	3.082	2.875	2.235	2.396	**13**
Inorganic P	Litter	0.012	0.010	0.021	0.007	0.006	0.003	**65**	0.003	0.001	0.011	0.004	0.001	**107**
	Topsoil	1.655	0.967	0.498	0.459	0.965	0.628	**52**	1.344	0.951	0.568	0.904	0.641	**35**
	Subsoil	7.718	7.345	4.460	2.929	8.116	5.582	**34**	7.732	7.680	4.883	6.696	5.762	**19**
	Total	9.385	8.322	4.979	3.395	9.086	6.214	**35**	9.078	8.631	5.462	7.605	6.404	**20**
Phosphate ions in soil solution	Litter	0.010	0.010	0.017	0.007	0.005	0.005	**52**	0.003	0.001	0.005	0.002	0.001	**63**
	Topsoil	0.018	0.015	0.008	0.016	0.012	0.003	**45**	0.008	0.004	0.012	0.003	0.002	**67**
	Subsoil	0.017	0.004	0.006	0.002	0.008	0.003	**83**	0.003	0.005	0.008	0.005	0.002	**44**
	Total	0.045	0.029	0.031	0.024	0.025	0.012	**40**	0.014	0.010	0.025	0.011	0.006	**54**

CV: coefficient of variation (ratio of the standard deviation to the mean, in %).

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Table 3. Continued.

P pool	Layer	Forest BAR	CHE	KRA	SAE	SAW	TOM	CV	Grassland BAR	CHE	KRA	SAE	TOM	CV
Diffusive phosphate ions in 1 day	Litter	0.003	0.002	0.003	0.002	0.001	0.000	**56**	0.000	0.000	0.001	0.001	0.000	**118**
	Topsoil	0.094	0.083	0.039	0.052	0.073	0.054	**32**	0.068	0.038	0.059	0.052	0.039	**25**
	Subsoil	0.397	0.480	0.447	0.110	0.741	0.532	**45**	0.280	0.464	0.496	0.187	0.588	**41**
	Total	0.494	0.565	0.490	0.164	0.815	0.586	**41**	0.348	0.503	0.557	0.240	0.627	**35**
Diffusive phosphate ions in 1 week	Litter	0.003	0.002	0.003	0.002	0.001	0.000	**56**	0.000	0.000	0.002	0.001	0.000	**122**
	Topsoil	0.148	0.144	0.064	0.081	0.136	0.108	**31**	0.106	0.064	0.101	0.093	0.077	**20**
	Subsoil	0.707	0.898	0.843	0.223	1.535	0.937	**49**	0.487	0.882	0.961	0.313	1.142	**46**
	Total	0.858	1.044	0.910	0.306	1.673	1.045	**45**	0.593	0.945	1.064	0.407	1.219	**40**
Diffusive phosphate ions in 3 months	Litter	0.003	0.002	0.003	0.002	0.001	0.000	**56**	0.000	0.000	0.002	0.001	0.000	**126**
	Topsoil	0.260	0.291	0.117	0.141	0.304	0.267	**35**	0.186	0.121	0.202	0.199	0.187	**19**
	Subsoil	1.506	2.037	1.951	0.563	3.979	2.032	**56**	1.005	2.045	2.286	0.616	2.623	**50**
	Total	1.769	2.330	2.071	0.705	4.284	2.300	**52**	1.191	2.166	2.491	0.816	2.810	**45**

CV: coefficient of variation (ratio of the standard deviation to the mean, in %).

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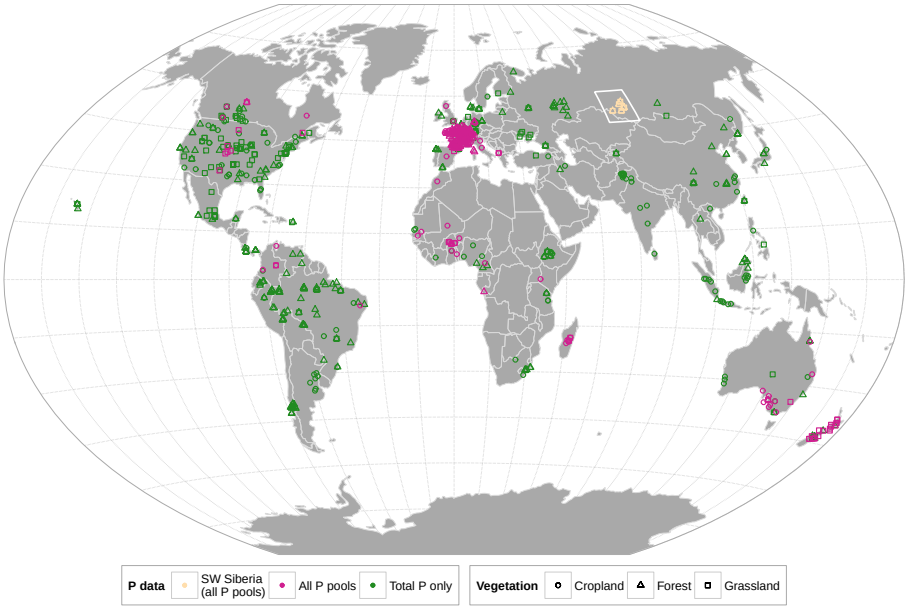

Figure 1. Localization of the study sites (south-western Siberia is highlighted in white, data points in beige) and of the data points from a literature compilation. Distinction is made between the quantity of information available for each point (color) and between vegetation covers (shape). "All P pools" stands for: total P, organic P, inorganic P, phosphate ions in soil solution, diffusive phosphate ions and isotopically exchangeable phosphate ions. Winkel-Tripel projection, graticules 15°.

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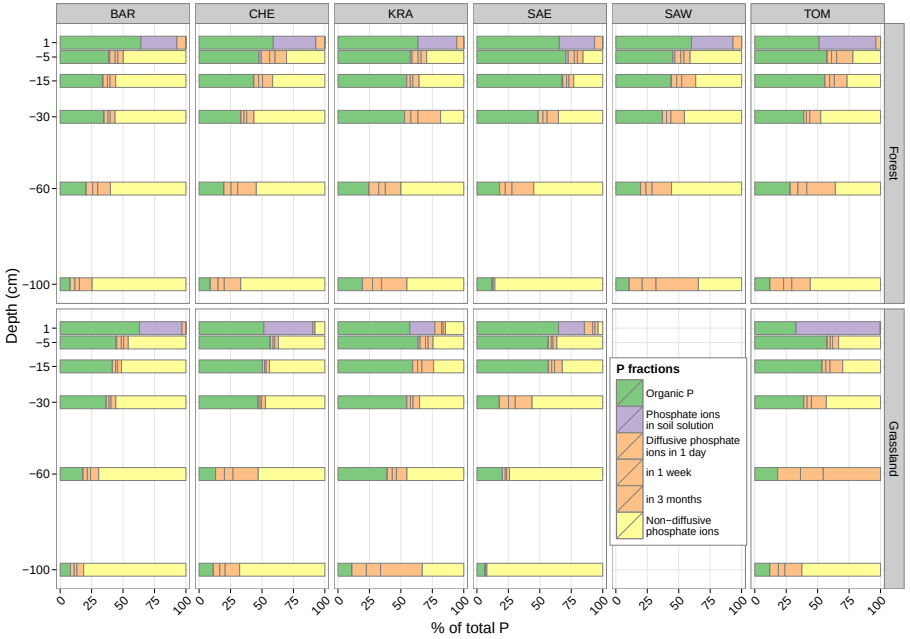

Figure 2. Structure of total P (P_{tot}) in terms of P ions in the soil solution (*Qw*), diffusive P ions at different time scales (*Pr*) and non-diffusive, or diffusive in more than 3 months, P ions ($P_{non\text{-}diff}$). Each fraction of P is expressed as % of P_{tot}. Note that the diffusive fractions at short times are included in the diffusive fractions at longer time (e.g. *Pr*(1 day) is included in *Pr*(1 week) and they are both included in *Pr*(3 months)). The depth "1" is the litter. "Litter" means all the dead plant material deposited on the soil surface (senesced leaf litterfall, small branches and senesced understorey vegetation in forests; senesced herbaceous vegetation in grasslands) that is to say mainly OL and OF horizons, and eventually OH (at BAR, CHE, KRA and SAE) at the date of sampling.

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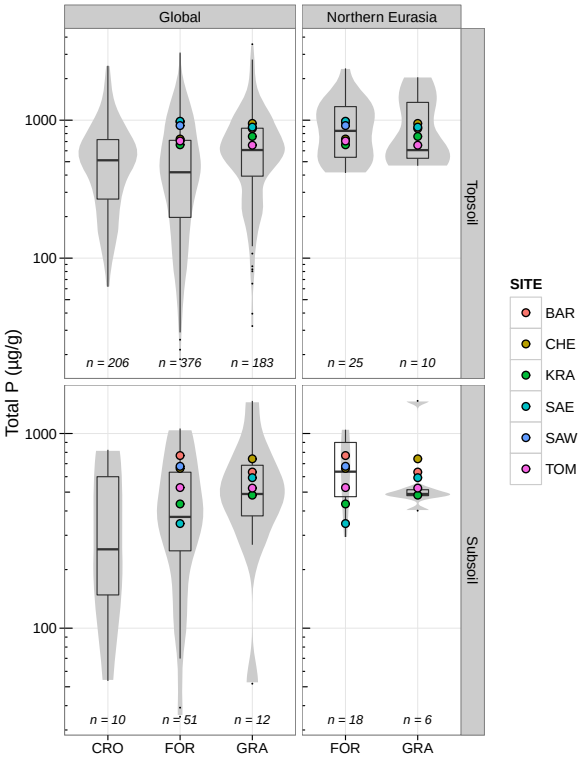

Figure 3. Comparison of total P concentrations in topsoils (about 0 to −20 cm) and subsoils (−20 to −100 cm depth) of south-western Siberia (colored dots) with similar vegetation contexts (CRO: croplands; FOR: forests; GRA: grasslands) at the global scale and at the Northern Eurasian scale (box- and violin-plots). The "*n*" provided indicates the number of individual points used to build the box- and the violin-plots.

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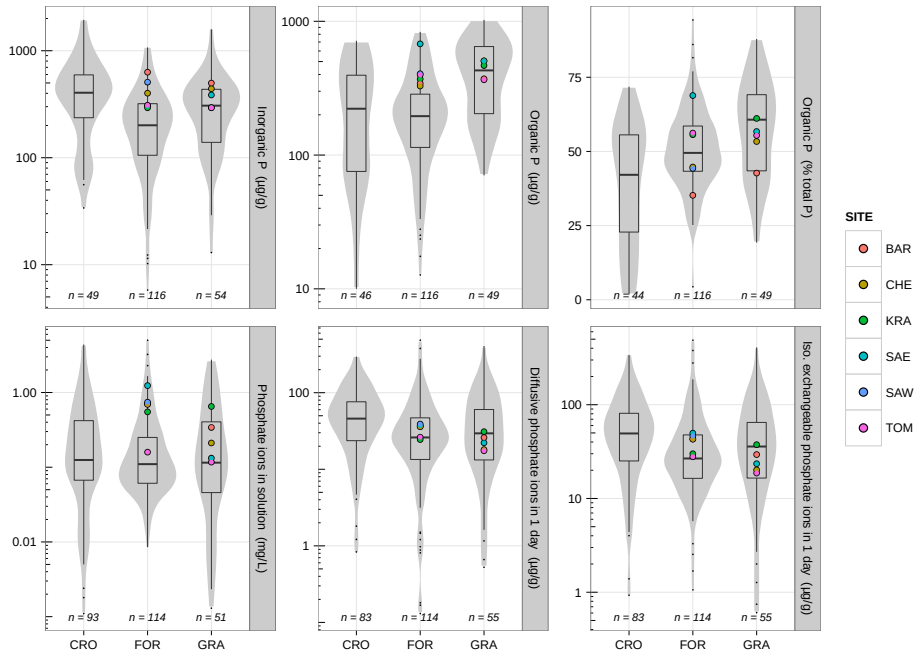

Figure 4. Comparison of the different components of the P status in topsoils (about 0 to −20 cm) of south-western Siberia (colored dots) and different vegetation types (CRO: croplands; FOR: forests; GRA: grasslands – box- and violin-plots) at the global scale (croplands and grasslands) or at the country scale (forests mostly located in France, along large gradients of soil and geology contexts). "Iso. exchangeable" is an abbreviation for "isotopically exchangeable". The "*n*" provided indicates the number of individual points used to build the box- and the violin-plots.

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