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Erosional effects on terrestrial resources over the last millennium in Reykjanes, southwest Iceland ☆

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ABSTRACT

The study presents the effect of soil erosion on vegetation, soil accumulation (SA), SA rate (SAR), soil quality, soil mass, and the soil organic carbon (SOC) pool in Brown Andosols and Histosols in a 24-km^2 area in southwest Iceland. Undisturbed prehistoric soils were distinguished from disturbed historic soils using tephrochronology. Soil erosion has been severe during historic time (last 1135 yr), resulting in the increase of the soil mass deposited in soils covered by vegetation by a factor of 7.3–9.2 and net loss of soil in unvegetated areas. The SAR correlated positively with SOC sequestration. SOC is easily transported and, given the extensive accumulation of soil, the net effect of burial and subsequent reduction in decomposition is to increase SOC storage. Nevertheless, the increased accumulation and soil depletion has decreased soil quality, including the SOC, and reduced soil resistance to erosion with the depleted SOC contributing to enrichment of atmospheric CO_2. The initial terrestrial disturbance was triggered by anthropogenic land use during the Medieval Warm Period, followed by volcanic activity approximately three centuries later. The combination of harsh climate during the Little Ice Age and drastic anthropogenic perturbations has led to land degradation at a catastrophic scale.

Introduction

The Icelandic environment has been strongly influenced by natural processes during the Holocene. Since settlement in AD 874, the introduction of grazing animals and other land use has drastically affected the natural environment. This includes the diminishing of the birch (*Betula pubescens*) woodlands and other vegetative cover (Hallsdóttir, 1987; Edwards et al., 2005; Erlendsson, 2007; Gathorne-Hardy et al., 2009), which has led to soil exposure and accelerated erosion over large areas, especially when in conjunction with harsh climate (e.g., Gísladóttir 1998; Arnalds et al., 2001; Ólafsdóttir and Guðmundsson, 2002; Gisladottir et al., 2005; Dugmore et al., 2009). This has specifically impacted processes and properties of volcanic soils (Andosols), which are subject to cryoturbation, landslide and accelerated erosion by wind and water (Wada, 1985; Arnalds, 2004, 2008).

While approximately 46% of the land surface in Iceland has sustained continuous vegetation cover, large areas have lost some or all of their soil cover formed during the postglacial era. Elsewhere, remaining soils have sparse or no vegetation cover (Gísladóttir, 1998; Guðjónsson and Gíslason, 1988; Arnalds et al., 2001), thus impairing soil carbon (C) sequestration. Among their multifunctional roles, soils support plant growth, increase soil biotic activity, enhance nutrient storage and strengthen the cycling of water and nutrients. In contrast, soil degradation by accelerated erosion and other processes impairs soil quality, reduces soil structure and depletes the soil organic matter (SOM) pool (e.g., Brady and Weil, 2002; Lal, 2003). Depletion of the SOM pool has also global implications because the terrestrial C pool is the third largest pool (Batjes, 1996) and strongly impacts the global C cycle (Lal, 2004).

Erosional–depositional processes affect soil organic C (SOC) distribution in two ways: it is depleted by erosion and increased by deposition. Some SOC-enriched sediments are redistributed over the landscape, while others are deposited in depression sites and transported into aquatic ecosystems (e.g., Stallard, 1998; Jacinthe and Lal, 2001; Smith et al., 2001; Lal, 2004; Rodríguez et al., 2004; Geirsdóttir et al., 2009; Gathorne-Hardy et al., 2009). SOC decomposition processes are severely constrained in some environmental settings and any SOC buried under anaerobic conditions is protected against decomposition. Yet, the impact of the SOC transported by erosional processes (e.g., Óskarsson et al., 2004) and redistributed over the landscape is not fully understood because the variability in its turnover characteristics has not been widely studied (Smith et al., 2001). Thus, the fate of C transported by erosional processes remains a debatable issue: is soil erosion a source or a sink of atmospheric CO_2? (e.g., Behre et al., 2007).

☆ Soil erosion and terrestrial resources.

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For erosion to be a net CO_2 sink, a full account of eroded C is required. Based on such assumptions, Stallard (1998), Harden et al. (1999), Yoo et al. (2005, 2006) and Van Oost et al. (2007) concluded that soil erosion is a global C sink. In contrast, other researchers have emphasized that soil erosion represents a source of atmospheric CO_2 (Gregorich et al., 1998; Jacinthe and Lal, 2001; Lal, 2003, 2004; Óskarsson et al., 2004; Polyakov and Lal, 2004; Rodríguez et al., 2004).

Regardless of being a source or sink, erosion strongly impacts soil quality on- and off-site (Lal, 1999). While the impacts of historic and contemporary erosion on the quality of volcanic soils of Iceland have not been widely studied, credible data on erosional impacts on soil quality are needed and are essential to land use planning for implementation of restorative and conservation-effective programs.

In Iceland, Óskarsson et al. (2004) mapped the extent of depositional soils in vegetated areas with high SOC content and areas with a severely depleted SOC pool. They suggested that 120–500 Tg of SOC had been removed by soil erosion over the past millennium, of which approximately 50% was oxidized and emitted into the atmosphere as CO_2. Over large river catchments in Iceland, active erosion and soil degradation have resulted in net SOC loss, with the remaining soils in the catchment not compensating for the large annual efflux (Kardjilov et al., 2006). Given the high C storage in Andosols and the fact that SOC does not decrease with depth (Batjes, 1996; Óskarsson et al., 2004; Peña-Ramírez et al., 2008), accelerated erosion of such soils may drastically increase C accumulation in depositional sites. Furthermore, the eroded sites would lose a considerable amount of SOC pool depending on the depth of soil truncated by erosion. Thus, there is strong need to study the effects of accelerated erosion on SOC pool in landscapes vulnerable to erosion–deposition cycles. Quantitative data of this nature does not exist for soils of Iceland, especially from studies in which the SOC pool is compared among undisturbed, uncultivated soils with those of perturbed by aeolian deposition.

Using multi-proxy evidence derived from the analyses of pollen and soil properties, the objective of this study was to determine the effects of soil erosion on the terrestrial environment, including vegetation, soil accumulation and the SOC pool during the late Holocene in the degraded environment of Krýsuvíkurheiði. The study, assisted by the presence of the *Landnám* tephra (AD 871 ± 2; Grönvold et al., 1995), was based on the assumption that the pre-settlement erosion and C cycling were at steady-state equilibrium. Field sampling and laboratory analysis, including pollen and soil properties, were designed to test the hypothesis that accelerated post-settlement erosion, induced or accelerated by anthropogenic and natural forces, resulted in the depletion of the ecosystem carbon pool and a decline in soil quality.

Study area and sites

The research site, Krýsuvíkurheiði, covers 24 km^2 of the Reykjanes peninsula, southwest Iceland (Fig. 1). The peninsula is characterized by postglacial lava covered by moss heath and sub-glacial hyloclastite ridges and mountains without substantial vegetative cover (Jóhannesson and Sæmundsson, 1988; Guðjónsson and Gíslason, 1988). Krýsuvíkurheiði was formed during the late Pleistocene and comprises gently sloping lava from shield volcano with striated lava indicative of interglacial formation (Jónsson 1978, Gísladóttir, 1998). The area was not covered by the Icelandic inland ice sheet in Younger Dryas time at ~11,300 ^{14}C yr BP (Norðdahl and Pétursson, 2005), signifying soil formation over the past 11,350 yr. Krýsuvíkurheiði is bounded by the Ögmundarhraun lava, formed AD 1151 to the west and the prehistoric postglacial, Eldborgarhraun to the east (Jóhannesson and Einarsson, 1988; Jónsson, 1998). Krýsuvíkurheiði is relatively flat (<100 m a.s.l.) and meets the North Atlantic Ocean in the south. The Reykjanes peninsula lacks major rivers and only two small brooks provide surface drainage in the study area.

The inhabitants of the study area practiced farming, based on animal grazing, from the 9th up to the mid-20th century. The initial settlement site (Fig. 1) was destroyed by the lava formed in AD 1151. Ruins of an indeterminate number of farms and a church are partly covered by lava over a 16-km^2 area (Jóhannesson and Einarsson, 1988). Based on historical documents Rafnsson (1982) suggested that there was a church in the area, indicating settlement from around AD 1200, although it declined during mid-16th century. The location is not known, but from the late 17th century the settlement was located farther east (Fig. 1) until it declined before the mid-20th century (Gísladóttir, 1998). Krýsuvíkurheiði was the largest and best pasture on the Reykjanes peninsula (Gísladóttir, 1998) and must have been used as grazing land by local farmers and outsiders following the eruption, as it is at present.

Krýsuvíkurheiði is characterized by severe soil erosion and highly degraded and susceptible landscapes (Fig. 1). About 20% is covered by dwarf shrub heath, 5% by grassland heath, 4% by grassland and 3% by wetland. Moss has been reestablished on 1% of the heavily degraded soils. Grasslands and wetlands are the least eroded plant communities. The remaining area is heavily degraded or barren and includes areas that have lost the entire soil cover (Gísladóttir, 1998, 2001). Predominant soils in the region are Brown Andosols (Arnalds and Grétarsson, 2001) with part of the area covered by Histosols (current study).

The climate is mild, cold-temperate oceanic. Between 1961 and 1990, the annual precipitation in the area was 1400–2000 mm, mean annual temperature was 4.7°C (maximum July–August 9.9°C and minimum in January 0.4°C) and on average 190 days per year exceeded 4°C. The average wind velocity (1952–2007) was 6.7 m s^{-1} and wind velocities below 4.4 m s^{-1} were rare. During cold periods, predominant winds blow from northerly directions (Einarsson, 1988; Icelandic Meteorological Office database, 2001, 2007).

Soil cores were obtained at four sites (Fig. 1): (i) Brown Andosol covered by dwarf shrub heath (K4), (ii) grassland heath (K7), (iii) grassland (K8), and (iv) Histosol dominated by sedges (K6).

Methods

Chronology

Tephra layers were identified visually within the soil cores. A sample from each tephra layer was also analyzed for major elemental composition to verify the literature-constructed tephra stratigraphy for the study area. Four of these samples were obtained from site K4 and one from K6 (Fig. 2). Determination of the origin of tephra layers was made on the basis of comparison of our results with cataloged reference data, which contain the geochemical signatures of tephra layers, unique to each volcanic system (e.g., Sigurgeirsson, 1992; Newton et al., 2007; Larsen and Eiríksson, 2008).

Pollen analysis

Sub-samples for pollen analysis (1 cm^3 each) from K6 were collected every two centimeters. The pollen preparation followed standard NaOH, HF, acetolysis methods (Moore et al., 1991) with the addition of *Lycopodium clavatum* spores (Stockmarr, 1971). Silicon oil of 12,500 cSt viscosity was the mounting medium. Pollen and spores were identified using the key of Moore et al. (1991). The separation of pollen between *Betula pubescens* from *Betula nana* was done by assigning all *Betula* pollen ≤20 µm to *B. nana* and the remainder to *B. pubescens* (Caseldine, 2001; Karlsdóttir et al., 2007). The terrestrial indigenous pollen count was ~250–300, excluding the predominant Cyperaceae, from which sums the percentages for all palynomorphs were calculated. Diagram construction used TILIA and TGView (Grimm, 1991, 2004). TILIA's statistical routine CONISS was used for numerical analysis and zonation of the pollen assemblages (Fig. 3). Pollen and spore taxonomy

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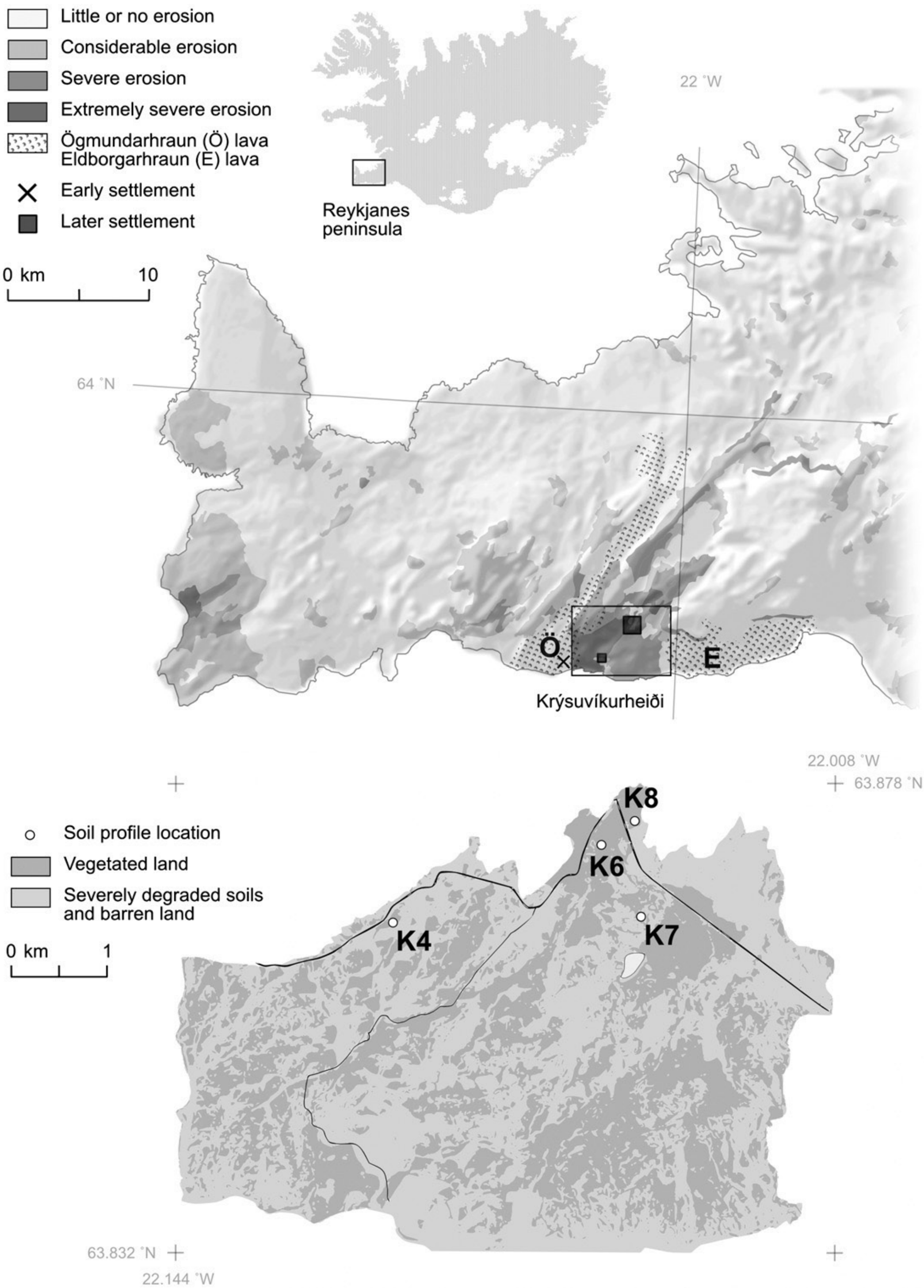

Figure 1. The study area, Krýsuvíkurheiði, on the Reykjanes peninsula (modified from Arnalds et al., 2001 and Gísladóttir, 1998).

follows the modified (Erlendsson, 2007) version of Bennett (2009). Plant nomenclature follows Kristinsson (1986).

Field and soil analysis

Field measurements were made for soil thickness through the profile to the youngest prehistoric tephra. Soil accumulation rates (SAR) (mm yr^{-1}) excluding tephra were calculated by measuring soil thickness between tephra layers of known age. While it is assumed that a constant SAR through the soil cores represents a stable environment, SAR is expected to vary with time during erosion phases. It is also assumed that if the dominant source material is of distant origin, it would result in small spatial SAR variation, whereas local and regional source material would result in larger variation (e.g., Dugmore et al., 2009).

The depositional structure of tephra layers was recorded since they provide information about the environmental situation at specific times. Horizontal tephra layers indicate no cryoturbation features, such as hummocks (*thufur*) which are related to heathland vegetation in drained soils. Hummocks are susceptible to erosion and have been formed at the expense of woodland and herb communities (Aradóttir et al., 1992; Gísladóttir, 2001; Ólafsdóttir and Guðmundsson, 2002).

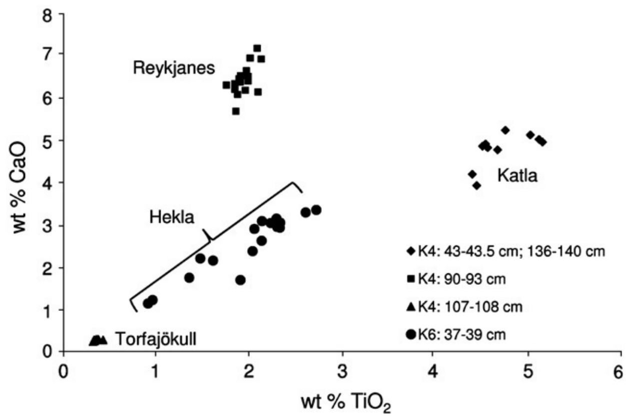

Figure 2. Results from the geochemical analysis of tephra shards. The homogeneity of the four different sources of tephra is indicated by using TiO_2 and CaO.

Soil cores were stored at field-moisture content at 4°C pending analysis. Soil samples were analyzed at a 3-cm interval for core K4 ($n=51$), every 2 cm for cores K7 ($n=31$) and K8 ($n=28$), and every 1 cm for K6 ($n=37$). If an abrupt textural change was observed, subsampling was adjusted accordingly. Soil bulk density (BD) was determined on a dry weight basis (Burt, 2004). Each sample was weighed moist, then air-dried in an oven at 60°C for 24 h, and passed through a 2-mm sieve to calculate the field-moisture content.

Prior to C and N analyses, samples were finely ground to pass through a 50-μm sieve. Total C and N concentrations were determined by the dry combustion method at high temperature (900°C) using a Vario Max C-N elementar analyzer. C and N concentrations were then used to compute the C:N ratio. Soil colour was measured on dry soil samples using a Minolta Chroma Meter CR-300. Soil pH was measured in water suspension (1:1) except for K6 (1:5), and in NaF solution (0.5 g soil in 30 mL) of 1 M NaF for estimation of andic properties of the soil following the method of Fields and Perrott as outlined by Blakemore et al. (1987). Synchronized measurements of BD, C concentration and soil layers enabled the calculation of C density (kg C m^{-2}). With concurrent soil dating, the SOC sequestration rate was computed (kg C m^{-2} yr^{-1}). The SOC enrichment ratio (SOC ER) was estimated from the prehistoric context.

Soil BD, soil thickness and soil dating enabled calculation of soil accumulation (kg m^{-2}) and soil enrichment ratio (SER) for different time periods during historic times. Soil degradation was assessed using the increased aeolian material in the soil (minerogenic material) and increased BD. The latter is related to minerogenic matter, either due to increased aeolian deposition or degradation of vegetation reducing the C sink function, or a combination of both. Reduced water cycling, indicated by reduced water content was measured on dry matter basis, decreased SOC, increased BD and soil accumulation rate resulting in decreased soil strength (Wada, 1985). Reduced pH (H_2O) is related to decrease in soil nutrients, because pH H_2O is positively related to cation exchange capacity (CEC), which is high for volcanic soils (Arnalds, 2008). An increase in the C:N ratio was used as an indicator for reduced N available to plants.

The question of whether historic erosion is a source or sink of atmospheric CO_2 was estimated from the mass balance of C in all cores from a specific time period, based on the tephra chronosequence, and by interpreting the mass balance in relation to the ecosystem C pool as affected by accelerated erosion.

Results

Chronostratigraphy

The results from the major elemental analyses show a clear distinction between the four different volcanic sources (Fig. 2). Three of the profiles have tephra-derived basal dates and all four contain the V (AD 871) tephra and top dates (surface/present). The tephra layers observed were Katla (K) tephra from ~1590 BC (Róbertsdóttir, 1992a), found in profile K4; Hekla (H) tephra from ~590 BC (Róbertsdóttir 1992b) found in K6 and K7; Vatnaöldur/Torfajökull (V; *Landnám*) tephra from AD 871±2 (Grönvold et al., 1995) found in all profiles; Reykjanes tephra (R; Medieval) tephra from AD 1226 (Jóhannesson and Einarsson, 1988) found in all profiles; Katla (K) tephra from ~AD 1500 (e.g., Hafliðason et al., 1992). Of the five tephras, the Medieval tephra is the most prominent. It forms a clear, coarse-grained stratigraphic marker, covering almost the entire Reykjanes area and beyond (Sigurgeirsson, 1995; Hafliðason et al., 1992) and provides the basis for an instant demarcation of sediments into pre- and post-AD 1226 parts. A scoria from the Ögmundarhraun lava from AD 1151 (Jóhannesson and Einarsson, 1988) was present in K4 and not in other profiles, which are outside the distribution of the scoria to the east. Minor differences observed in the tephra stratigraphy among the sites are normal because of taphonomic factors (e.g., preservation of tephra shards and different depositional conditions in existence at any given time at each sampling site can vary).

The tephras can also be used to define broader environmentally significant phases, as they would seem to coincide with major post-settlement climatic changes. While the V (AD 871) tephra caps a stratigraphic record for pristine environments, its time of deposition approximates the onset of the Medieval Warm Period (MWP). Although the time of the MWP–LIA transition is somewhat ambiguous, evidence for climatic deterioration from the mid-13th century is inferred from marine and lacusterine records with increasingly harsh conditions from around AD 1500 (e.g., Eiríksson et al., 2006; Sicre et al., 2008; Gathorne-Hardy et al., 2009).

Vegetation changes

On the basis of CONISS and the visual inspection of the data, the pollen record can be divided into three local pollen assemblage zones (LPAZ) (Table 1; Fig. 3). The first LPAZ (K6-I; 590 BC–AD 1370; 37–22 cm) is characterized by the pollen of woody taxa, although at relatively low values (*Betula* sp. [<10% TLP], and *Salix* [<5% TLP]), the damp-loving and grazing-sensitive tall herbs *Filipendula ulmaria* and *Angelica* sp. (both taxa typically 5–15% TLP) and some taxa typical for open environments (e.g., *Thalictrum alpinum*, *Rumex*-type, Poaceae and Cyperaceae). In LPAZ K6-II (AD 1370–AD 1600; 22–14 cm) there is an increased signal for *B. pubescens* (from <10% to 10–23% TLP) whereas the tall herbs are reduced to negligent values, probably because of grazing. There is also a rise in *T. alpinum*, a common inhabitant of pasture communities, which typically expands after settlement in Icelandic pollen diagrams (e.g., Hallsdóttir, 1987; Erlendsson et al., 2009a). The third LPAZ (K6-III; AD 1600–AD 2006; 14–0 cm) reflects an environment affected by a full-scale human impact, attested by the almost complete disappearance of *B. pubescens* and a rise in taxa that benefit from agricultural activities such as Poaceae, *Selaginella selaginoides*, *Plantago maritima* and Lactuceae.

Soil erosion, soil quality and SOC sequestration

The age resolution of the soil samples is centennial from the prehistoric to AD 1226 and decadal thereafter. However, K4 has a resolution on decadal scale for the entire historic time. Soil profile K8 lacks prehistoric tephra, which prevents comparison between historic and prehistoric soil properties. Similarly AD 1500 tephra is lacking in soil profile K6.

Total soil thickness varies among the soil cores (Fig. 4), being greatest in K4 and the least in the Histosol (K6). There is a large difference in the soil accumulation during historic and prehistoric time (Table 2). Since 590 BC, 86–88% of the soil density (kg m^{-2}) (K4, K6, K7) was formed during the historic time. The impact of historic soil accumulation is further emphasized by the high SER.

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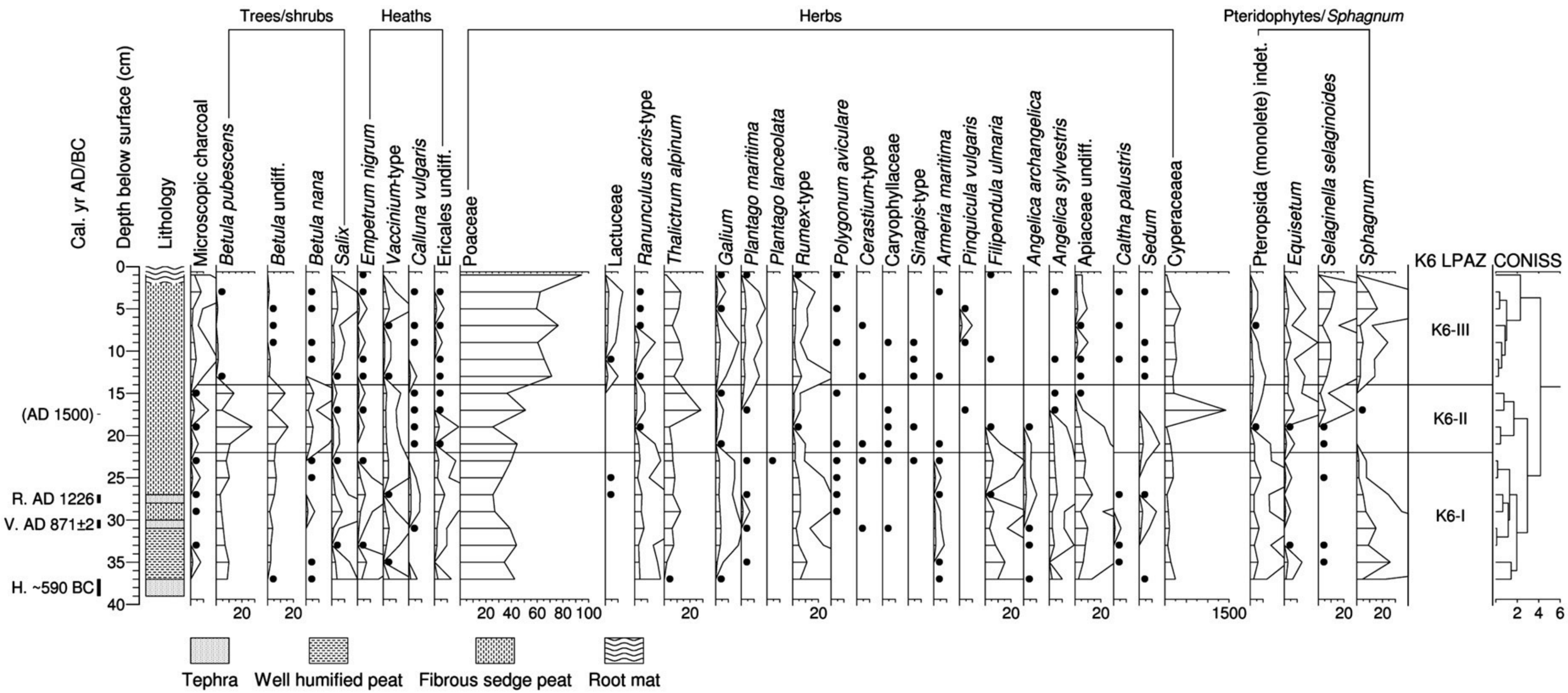

Figure 3. Pollen percentage diagram from site K6. Selected taxa are displayed. Exaggeration curves are ×5. Date in bracket denotes extrapolated age. Dots signify values below 1% TLP.

Table 1
Major characteristics of the local pollen assemblage zones for profile K6.

LPAZ	Depth (cm)	Age (AD/BC)	Characteristic and main advancing taxa from previous zone	Main retreating taxa from previous zone
K6-III	14–0	AD 1600[a]–AD 2006	*Salix*, Poaceae, Lactuceae, *Galium*, *Plantago maritima*, *Rumex*-type, *Sphagnum*	*Betula* sp., *Ranunculus acris*-type
K6-II	22–14	AD 1370[a]–AD 1600[a]	*Betula* sp., Poaceae, *Thalictrum alpinum*, Cypearceae, *Equisetum*, *Selaginella selaginoides*	*Salix*, heaths, *Filipendula ulmaria*, Apiaceae sp., *Sphagnum*, *Rumex*-type, *Galium*
K6-I	37–22	590 BC–AD 1370[a]	*Betula pubescens*, *Salix*, heaths, Poaceae, *Ranunclulus acris*-type, *Thalictrum alpinum*, *Rumex*-type, *Filipendula ulmaria*, Apiaceae sp., Cyperaceae, *Sphagnum*	N/A

[a] Extrapolated age.

The amount of C sequestrated in the soil is influenced by the total amount of soil.

The SAR correlates positively with the C sequestration rate in the Brown Andosols ($r = 0.949$; $p < 0.0001$) and for the combined Brown Andosols and Histosols ($r = 0.851$; $p < 0.0001$). The impact of SA on the SOC density is evident by the increase in SOC density during historic time. As much as 78–85% of the SOC density accumulated since 590 BC was formed during the historic time in the Brown Andosols and the Histosol (Table 3; Fig. 5). The historic SOC density is greatest in K4 (27.5 kg m^{-2}), followed by that in K6 (21 kg m^{-2}), while the historic SOC densities are similar in K7 and K8 (17 kg m^{-2}). The impact of soil erosion on the historic SOC budget is evident by the large SOC ER estimated from the prehistoric context (Table 3).

Soil properties before settlement (ca AD 871)

The data for prehistoric soil properties indicate a stable environment, reflected by good soil quality, high SOC concentration, and relatively high pH values in the Brown Andosols and low pH, low BD, and high moisture content in the Histosols (Fig. 6). The pH NaF values, ranging from 10.4 to 11.7 indicate the andic properties of the soils. The SAR is low, and ranges from 0.05 mm yr^{-1} in K7, 0.1 mm yr^{-1} in K4, and 0.04 mm yr^{-1} in the Histosols (Fig. 4). The Histosol is very dark (9YR 3.7/2.1–8.5YR 3.4/2), and comprises highly humified and decomposed peat. All soils become lighter in colour above the AD 871 tephra since the Norse colonization of the area. The annual SOC sequestration rate from 590 BC to AD 871 varies from 2.5 g C m^{-2} yr^{-1}, in K6 to ~3.2 g C m^{-2} yr^{-1} in K7 and K4 (Fig. 5). The prehistoric tephras show no indication of disturbance by cryoturbation.

Soil properties from AD 871 to 1226

Between AD 871–1226 (roughly the MWP), the change in soil properties indicates some ecosystem disturbance. This is especially reflected by increased aeolian material with the attendant increase in SAR and SER (Fig. 4; Table 2) in all cores, but most significantly in K4 with an increase in SAR by a factor of 3.3 (0.32 mm yr^{-1}). The SAR almost doubles during the succeeding period between AD 1151 and

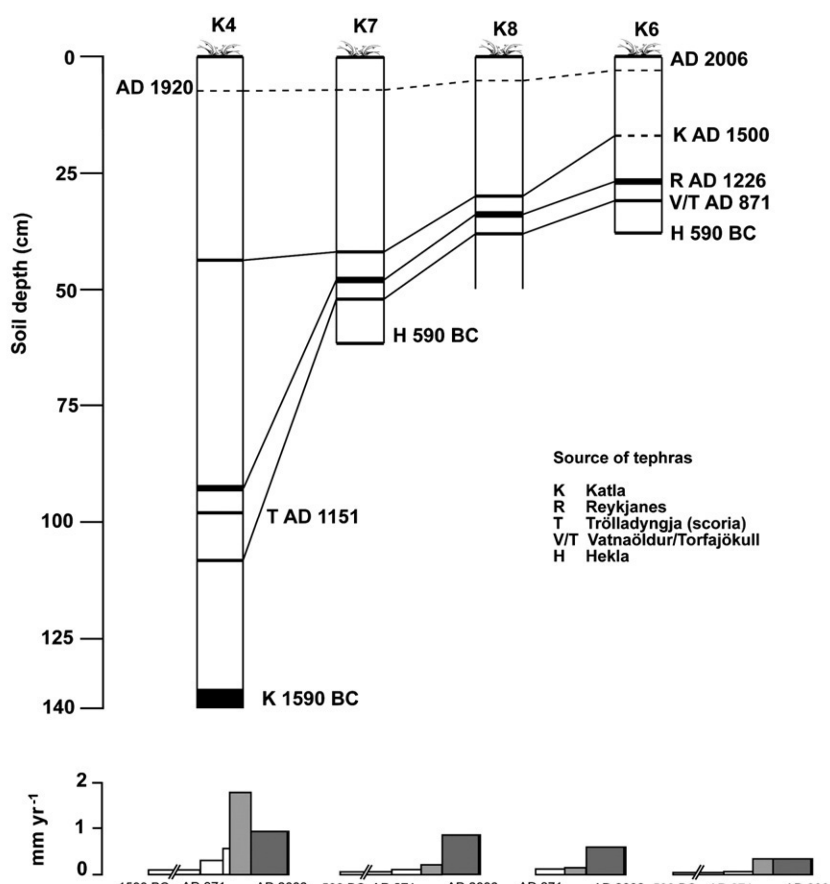

Figure 4. Soil depth (cm) and accumulation between tephras of known age (mm yr^{-1}). Soil accumulation rate is shown for different time periods. K4, K7 and K8 represent Brown Andosol and K6, Histosol.

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Table 2
Temporal changes in soil mass (kg m^{-2}).

Time Periods	K4 (n=51)			K7 (n=31)			K6 (n=31)		
	Soil	Postulated soil[a]	SER[b]	Soil	Postulated soil[a]	SER[b]	Soil	Postulated soil[a]	SER[b]
Historic (AD)									
1920–2006	26.8	5.4	5	25.3	2.4	10.5	9.3	0.8	11.1
1500–1920	245.8	26.2	9.4	165.4	12	13.8	38.7	4.1	9.4
1226–1500	299.1	17.1	17.5	28.8	7.8	3.7	29.9	2.7	11.1
871–1226	81.7	22.1	3.7	15.5	10.1	1.5	8.9	3.5	2.5
Total	653.4	70.8	9.2	235	32.3	7.3	86.8	11.1	7.8
Prehistoric (BC–AD)									
590–871				37.9			14.3		
1590–871	153.3								
Total	153.3			37.9			14.3		

[a] Soil accumulation postulated from the prehistoric contexts.
[b] Soil enrichment ratio.

AD 1226. The aeolian soil mass increased by 231 g m^{-2} yr^{-1} in K4 over the 355 yr in comparison to 62.3 g m^{-2} yr^{-1} during the preceding ~2460 prehistoric years, an increase in SER by a factor of 3.7 (Table 2). The colour changes from 9.3YR 4.9/3.7 during prehistoric time to 9.3YR 5/4.2 during AD 871–1226. The Histosol shows a pattern similar to that of K4, with increase in SAR by a factor of 2 and that in SER by a factor of 2.5, involving 25.1 g m^{-2} yr^{-1} during the first 355 yr of Norse settlement, compared to the 9.8 g m^{-2} yr^{-1} during the preceding period. The Histosol was very wet, as is indicated by the high values for moisture-loving taxa and the limited woodland cover, and was able to function as a sink of the potential aeolian material. The large aeolian deposition increased the SOC sequestration rate. This results in higher SOC density in the K4 Brown Andosols than in the Histosols (Table 3, Fig. 5).

The SOC sequestration rate was 13 g SOC m^{-2} yr^{-1} in K4, compared with 6.2 g SOC m^{-2} yr^{-1} in K6. The SOC sequestration in K7 and K8 was merely 25–32% of that in K4 (Fig. 5). A sharp decline in soil quality occurred immediately after ~AD 871, except in K4 where the soil quality declined after the eruption in AD 1151 (Fig. 6).

Reduction in SOC concentration, increase in the C:N ratio, increase in BD and decrease in water content, indicate a decline in soil quality (Fig. 6). The AD 871 tephra is horizontal but discontinuous. It is very thin (<5 mm) but a broken horizon might indicate either an exposed soil when the tephra was deposited or later erosion. The AD 1226 tephra is horizontal in all cores except K4, where the deposition structure is slightly undulating, indicating hummocks at the time of deposition.

Soil properties from AD 1226 to 1500

The environment during AD 1226–1500 was very unstable, as represented by the SAR in K4 where it reaches its highest level. The average SAR in K4 increases by a factor of 4.2 (1.79 mm yr^{-1}) (Fig. 4). Around AD 1340 (~70 cm depth) in K4, the soil becomes sandier and looser in consistency. There are thin lenses of sand above this level in the core, indicating diminished resistance to erosion in the surrounding landscape. The SAR in K7 and K8 increases by a factor of 1.9 and 1.3, respectively, even though the SAR is considerably less than in K4 and K6. The large increase in SAR signifies a large source of aeolian material both inside and outside the area. The aeolian deposition increased SOC, which is densest in K4 at 10.4 kg C m^{-2} and in K6 at 8.9 kg C m^{-2}. The SOC sequestration rate in the same cores was 38 and 32.5 g C m^{-2} yr^{-1}. The rate of SOC sequestration in K7 and K8 was less than 26% of that of K4 and K6 during this time.

Soil degradation is indicated by the decline in SOC concentration, increase in BD, decrease in water content and pH, and increase in C:N ratio (Fig. 6). Thus the soils were physically, chemically and biologically degraded.

The Histosol was still very wet until about AD 1400 (~21cm depth) (Fig. 6) and in combination with herbaceous species (Fig. 3) able to trap aeolian deposits. The excessive accumulation of soil in the Histosol is reflected by the change in colour from 8.5YR 3.5/2.1 to 9YR 4.4/2.8 during the period when the SAR increased from 0.08 mm yr^{-1} to 0.4 mm yr^{-1}.

Soil properties from AD 1500 to present

Accelerated soil erosion continued during AD 1500–1920 (often linked with the cold period of the LIA), as indicated by a lower spatial variation in the SAR. The Brown Andosol close to the K7 and K8 apparently accumulated less aeolian materials than K4. However, this changes after AD 1500, when the SAR increases by a factor of 4 in K7 and K8, but decreases in K4 (Fig. 4 and Table 2). The soil in K7 and K8 is enriched with aeolian deposits beginning at the end of the 16th century and mid-17th century onwards, when the soil becomes coarser and looser. The SOC rate in the Histosol decreases more

Table 3
Temporal changes in soil organic carbon (SOC) pool (kg m^{-2}).

Time Periods	K4 (n=51)			K7 (n=31)			K6 (n=31)		
	SOC	Postulated SOC[a]	SOC ER[b]	SOC	Postulated SOC[a]	SOC ER[b]	SOC	Postulated SOC[a]	SOC ER[b]
Historic (AD)									
1920–2006	3.1	0.3	11.3	2.8	0.3	10	1.2	0.2	5.5
1500–1920	9.4	1.3	7	11.4	1.4	8.2	8.5	1.0	8.1
1226–1500	10.4	0.9	11.8	2.3	0.9	2.5	8.9	0.7	13
871–1226	4.6	1.1	4.2	1.1	1.2	0.6	2.2	0.9	2.5
Total	27.5	3.6	7.6	17.6	3.8	4.7	20.8	2.8	7.3
Prehistoric (BC–AD)									
590–871				4.7			3.6		
1590–871	7.9								
Total	7.9			4.7			3.6		

[a] Soil organic accumulation postulated from the prehistoric contexts.
[b] Soil organic carbon enrichment ratio.

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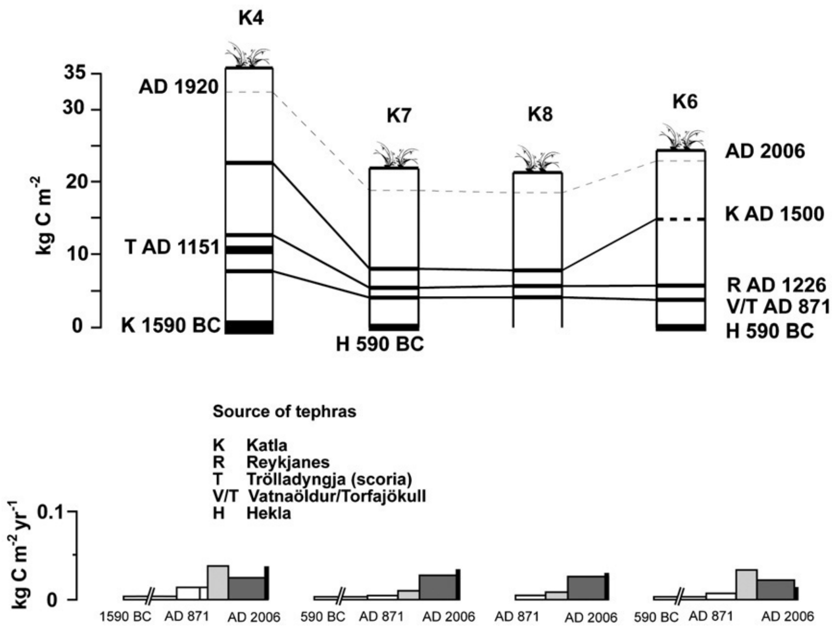

Figure 5. Soil organic carbon sequestration (kg C m^{-2}) and annual organic carbon sequestration rate (kg C m^{-2} yr^{-1}) for the different soil sites.

than does the SAR. The AD 1500 tephra forms a discontinuous layer in K7, indicating that the soil was partly exposed when the tephra layer was formed or became eroded soon thereafter. Changes in SOC density are in accord with the amount of accumulated soil and the increase in BD, (Table 3; Figs. 5 and 6). Despite the high SOC density and SOC sequestration rate of 25–28 g m^{-2} yr^{-1} in the Brown Andosol and 19 g m^{-2} yr^{-1} in the Histosol, there was an overall depletion of the SOC pool with an attendant decline in soil quality until the onset of recovery, in the late 19th century. Despite its continued recovery during the 20th century, soil quality has not yet attained the value similar to those of the prehistoric undisturbed soils.

Discussion

Vegetation

Vegetation data for the period ca. 590 BC–AD 1370 (Fig. 3; LPAZ K6-I; 37–22 cm) suggest that *B. pubescens* was not abundant around the sampling site. Even with Cyperaceae excluded from the TLP, the taxon does not reach above 10% TLP. In Hallsdóttir's (1987) pre-settlement records, where sedges are excluded from the TLP, *Betula* typically reaches values of 50–80% TLP. Strong signals for damp-loving taxa (Cyperaceae, *F. ulmaria*, Apiaceae sp.) indicate that the bog was reasonably wet. In contrast to many other sites that were supposedly wooded at settlement time (Hallsdóttir, 1987; Edwards et al., 2005; Erlendsson et al., 2009b), the data also show limited signal for environmental change around AD 871. The prolonged presence of grazing-intolerant taxa (e.g., *F. ulmaria* and *Angelica* sp.), which thrived until ca. AD 1400 (~15 cm), indicates that the surrounding environment was not grazed intensively and that permanent settlement had not been established near K6 by AD 1370. Such belated near-site settlement has also been suggested for Ketilsstaðir in south Iceland on the basis of pollen and Coleoptera. Like site K6, the Ketilsstaðir area was largely unwooded by AD 871 (Erlendsson et al., 2009a). The apparent late settlement around K6 is in agreement with the historical evidence and archaeological records that show the early settlement to be some >5 km to the southwest from the K6 site (Fig. 1; Rafnsson, 1982; Jóhannesson and Einarsson, 1988).

The pollen assemblages for ca. AD 1370–1600 (Fig. 3; LPAZ K6-II; 22–14 cm) reflect a period of environmental disturbance, perhaps caused by the introduction of near-site farming activities. This trend is indicated by the disappearance of the most grazing-sensitive herbs (*F. ulmaria* and *Angelica* spp.) and a significant increase in apophytic taxa (e.g., *T. alpinum*, *P. maritima*, *Equisetum*, *S. selaginoides*). A similar trend is also observed in Icelandic pollen diagrams, which signifies the initial settlement phase (Hallsdóttir, 1987; Erlendsson, 2007). Conversely, *B. pubescens* reaches its highest scores during this period.

This trend can be explained by two different scenarios: (i) as a consequence of an increase in soil erosion and the reworking of *Betula* pollen into the bog, or (ii) as a consequence of an expansion of woodland or increased flowering. Both scenarios could result from the thinning out of woodland around the bog. The latter scenario would seem the more likely as less dense woodland could allow the remaining trees to flourish, produce more pollen and decrease the filtering effects that dense woodlands can impose on pollen dispersal (cf. Lawson et al., 2007). This is also supported by the soil data, which indicate degradation attributed to an increase in BD, a decrease in moisture content, and a decline in SOC concentration (Fig. 6). However, values for *B. pubescens* plummet to very low values across the next zone boundary, at a time when soil erosion is at maximum in the surrounding landscape (e.g., sites K7 and K8 in Fig. 4). It is difficult ascertain whether the bog surface became drier during this period as the decrease in wetland taxa (e.g., *F. ulmaria*, *Angelica* sp. and *Sphagnum*) could also relate to the grazing and trampling of herbivores. If the bog became drier, as a result of an increased input of soil shown by the sedimentary proxies (Fig. 4) or by deliberate draining, this could also have facilitated the invasion of *Betula* sp. across the bog and be responsible for the increased values for birch pollen.

The most noticeable aspect in the pollen assemblages in LPAZ K6-III (AD 1600–2006; 14–0 cm) is the dwindling of *B. pubescens* as opposed to the expansion in Poaceae. This pattern is typical for Icelandic pollen diagrams (Hallsdóttir, 1987; Erlendsson et al., 2009b) and is interpreted as a reflection of the expansion of open grasslands following the decline of woodlands resulting from full-scale human impact. It is also likely the bog became increasingly wet from AD 1600 onwards, as is evident from the higher values for Cyperaceae and *Sphagnum* along with changes in soil properties. Such trends may be linked to a rising water table and wetness during the LIA and is also supported by a record of the increased frequency of sea ice off the coast of Iceland during 17th–19th century (Ogilvie and Jónsdóttir, 2000). The pollen data correspond well with the historical record in

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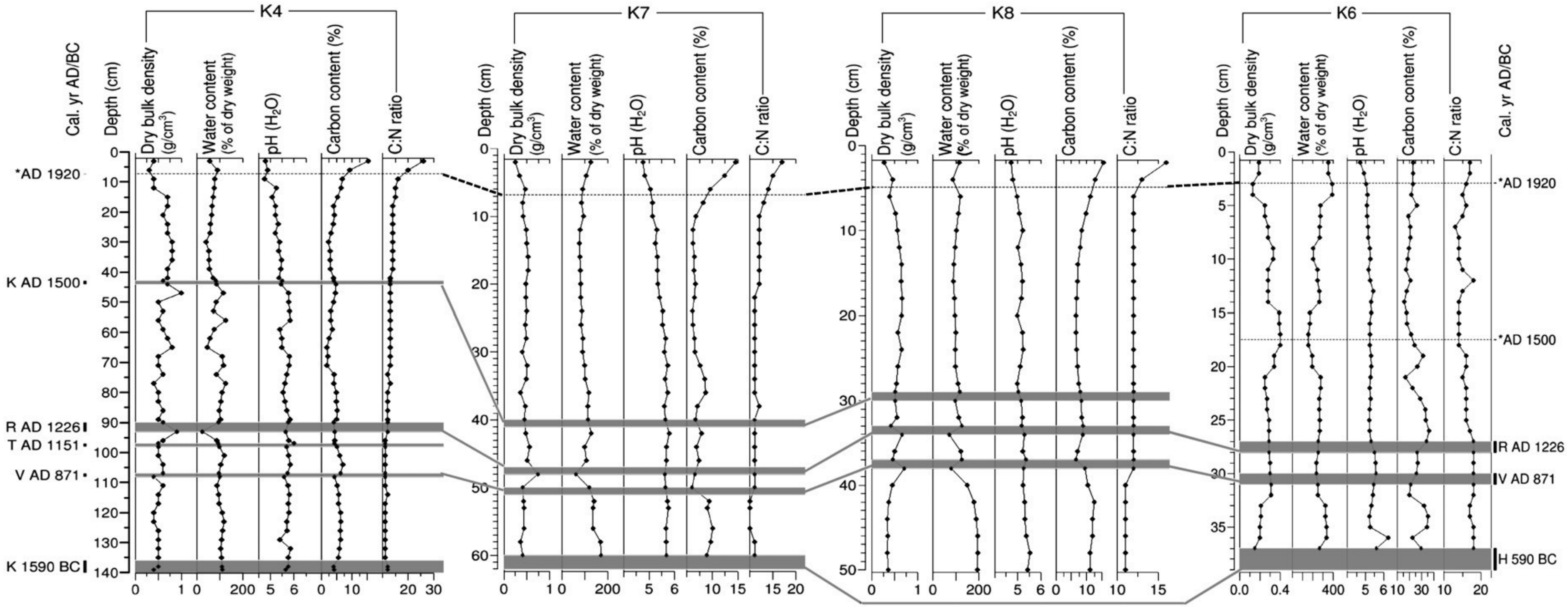

Figure 6. Vertical distribution of soil properties for the different soil sites. The *y*-axes scales vary.

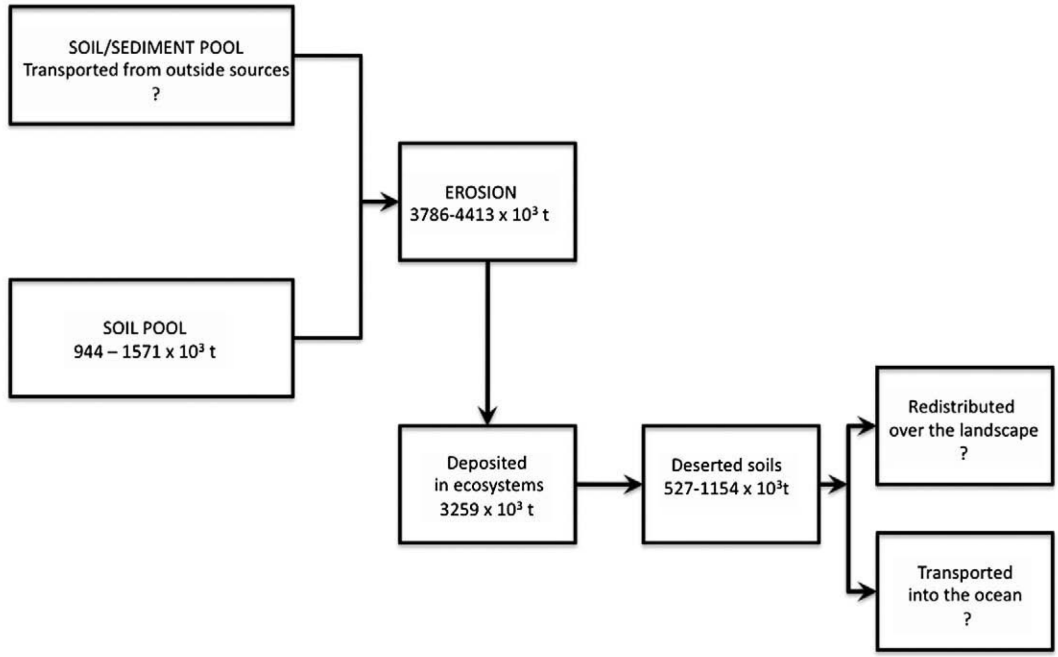

Figure 7. Mass balance of soil transported, deposited and redistributed over the landscape since settlement.

the sense that settlement at the modern Krýsuvík, near-site K6, was only established as late as in the late 17th century (Gísladóttir, 1998), which appears to roughly coincide with the time of birch decline. It is possible that the slightly earlier reduction in other, more vulnerable, grazing-intolerant taxa might be the result of increased grazing pressure, possibly related to diminished grazing areas in the wider region as a result of soil erosion. As noted above, Krýsuvíkurheiði is now the best pasture in the Reykjanes peninsula and is likely to have been exploited as such. Perhaps the area around K6 was initially used as a shieling site (grazing outpost) which later turned into a full farm.

Soil erosion

The increased SAR and SE in all profiles (Fig. 4; Table 2) during historical times accentuate the aeolian activity following the settlement, as is also observed elsewhere in Iceland (Þórarinsson, 1961; Dugmore et al., 2009). The spatial and temporal variation in SAR and

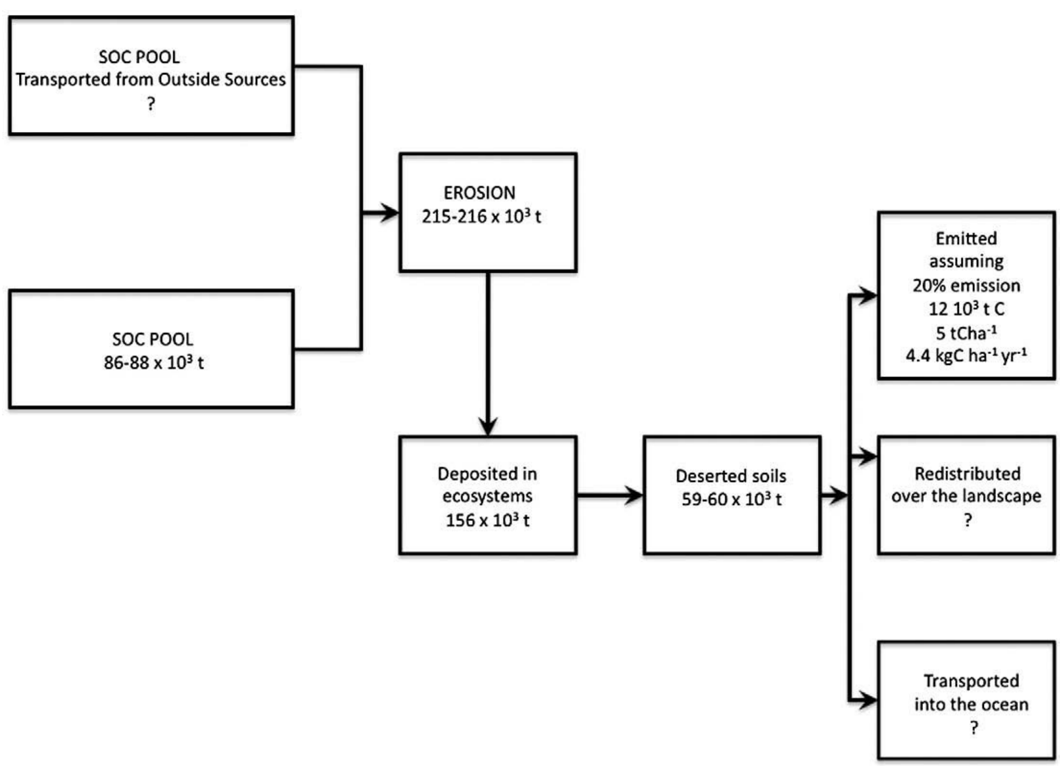

Figure 8. Fate of soil organic carbon transported, redistributed and emitted by erosional processes since settlement. Postscript: Emission assumed as done by Lal (2003). For soils of Iceland Óskarsson et al. (2004) assumed a rate of 50%.

SE during early historic time indicates that the source area is of local or regional origin (cf. Dugmore et al., 2009), especially in K4, reinforcing the idea that erosion started in the western part of the study area and later expanded to the eastern part. However, the large SAR and SE in the eastern wetland areas should not be neglected, because of the sink potential of the wetlands. It is likely that these wetlands consumed dispersed aeolian material transported from sources farther away from the immediate site, in contrast to the drained plant communities lacking shrub or woodland cover (K7 and K8) by that time (cf. McTainsh and Strong, 2007).

The deposition of the R-tephra added a wind-transportable and abrasive element to the already damaged vegetation cover in Krýsuvíkurheiði. The negative perception of the inhabitants of Reykjanes towards this tephra is reflected in naming the winter of AD 1226–1227 *sandvetur* (sand-winter; Jóhannesson and Einarsson, 1988). The R-tephra, cold spells during the 13th century AD (Dawson et al., 2007) and land use intensification because of loss of pasture associated with the formation of the 16 km^2 Ögmundarhraun lava conspired to mount the pressure on an ecosystem already strained by the preceding ~300 yrs of agricultural activity. The land use surpassed the carrying capacity of the ecosystem before AD 1500, after which the source of eroded material seems depleted, as inferred from decreased SAR and finer soil texture.

It is not after AD 1500 that the erosion culminates in the western part of the study area, at a time when the local aeolian sources close to site K4 seem to have been greatly reduced. Land degradation during LIA was further exacerbated by the anthropogenic impacts (Gísladóttir, 1998), as is also shown by the change in vegetation (Fig. 3), soil texture and a weaker soil structure. Moreover, the period after AD 1500, typically associated with the LIA, represents the most severe historical erosion era in Iceland (e.g., Gísladóttir, 1998; Ólafsdóttir and Guðmundsson, 2002; Geirsdóttir et al., 2009; Gathorne-Hardy et al., 2009), with the potential of long-distant aeolian transportation.

Given the heavily degraded soil and the present rate of erosion, soil degradation during historic time may have been a source for local aeolian material that was accumulated in the study area, especially in the vegetated sites, relocated over the landscape, and eventually transported to the ocean. To quantify the process, a soil mass balance after settlement in Krýsuvíkurheiði was computed (Fig. 7). It is assumed that by the time of settlement the area was fully vegetated and prehistoric soils represented stable environmental conditions (postulated soil). Presently barren land was assumed covered by soils of similar mass as prehistoric soils in K4 and K7 (postulated covered by heathland vegetation). Soil mass is based on soil data related to the dominating plant communities by soil cores and extrapolated to area coverage (cf. Gísladóttir, 1998). These values (Fig. 7) indicate that a considerable mass of soil was transported over the landscape. A minimum of 3786–4413 Gg of soil was displaced from local and regional sources, of which 3259 Gg was deposited on the present vegetated land. This was added to the 417 Gg of postulated soil formation, an increase by a factor of nearly eight. Significant mass has been lost and most likely a large part was transported into the ocean and/or redistributed over the landscape. These components are, however, difficult to estimate from the present data. The soil disturbances show clearly that even though Andosols in an undisturbed state are relatively resistant to erosion, the disturbed soils are extremely vulnerable to disturbance (cf. Wada, 1985).

Soil quality and soil organic carbon

The movement of the soil mass severely reduced soil quality, as has also been reported by others (e.g., Brady and Weil, 2002; Lal, 2003). There is a general decrease in soil quality during the historic time. Decline in soil quality is characterized by diminished water cycling, a rise in acidity with an attendant decrease in CEC (Arnalds, 2008), and an increase in C:N ratio with a possible decrease in N available to plants due to constant recharge of the ecosystem by aeolian deposition. Of primary importance is the conclusion that erosion depleted the SOC pool. The prehistoric value of SOC is higher than the historic level, except after AD 1920 (Fig. 6). The SOC concentration is the highest in K6, as is expected for a Histosol because of low decomposition under anaerobic conditions. The SOC concentration decreased below 20% after AD 1500. The higher temporal resolution of the data shows that SOC enrichment remains the same (0.030 kg C m^{-2} yr^{-1}) during the intervals AD 1226–1370 and AD 1370–1600. This trend suggests an increase in decomposition of SOC during the drier environmental conditions between AD 1370 and 1600 (Figs. 5 and 6) with a decrease to 0.017 kg C m^{-2} yr^{-1} between AD 1600 and 1920. The relatively larger SOC sequestration may also be related to the increase in coverage of shrubs and trees (Fig. 3; Jobbágy and Jackson, 2000) and/or an influx of organic-rich sediments. In addition to the severely degraded environment, severe cold was characteristic during this period (Ogilvie and Jónsdóttir, 2000; Eiríksson et al., 2006), which reduced plant production and adversely affected the SOC storage. Even though the soil quality of the Histosol decreases during historic time, the Histosols do not dry out as rapidly as do the Brown Andosols, indicating their stronger resistance to erosional processes.

To understand the fate of C over historic time (Fig. 8) for the Krýsuvíkurheiði region, soil data and present vegetation communities and coverage were used to compute the mass balance of the soil. It is apparent that the displaced SOC (215–216 Gg) by erosion is larger than the historic undisturbed SOC pool in Krýsuvíkurheiði (86–88 Gg). The present vegetated land is assumed to have sequestered 27 Gg of SOC under undisturbed conditions, whereas over 1135 yr 156 Gg have been added to that amount through deposition, an increase by a factor of more than 5. This increase in SOC is opposite to the trend in soil quality that has been depleted in the same soils. The deserted soils have lost 59–60 Gg C over the last 1135 yr. Contrary to the hypothesis of Stallard (1998), SOC has not been replaced in these deserted sites.

Given the positive correlation between soil depth and C sequestration, the majority of the deposited SOC (off-site) is derived from eroded soils (on-site). The low percentage of C in the reworked soils compared to undisturbed soils *in situ*, supports the contention that some of the SOC has been oxidized following erosion/deposition pathways (cf. Gregorich et al., 1998; Óskarsson et al., 2004; Yoo et al., 2005, 2006). Lack of vegetation at eroding sites reduces the soil's capacity to store C and to compensate for the loss of SOC. SOC moved from the exposed soils may be transported into the ocean, redistributed over the landscape and emitted into the atmosphere through decomposition. A considerable amount is assumed to have been transported into the ocean given the location of the site, with some redistributed over the area. Assuming that 20% of the SOC is emitted (Lal, 2003), this is equivalent to 4.4 kg C ha^{-1} yr^{-1} for the entire study area. Furthermore, assuming that 20% of the displaced SOC that has been added to ecosystem is also emitted to the atmosphere (Lal, 2003, 2004), this process would increase the erosion-caused emission by an additional 31 Gg C. This would equal to 11.4 kg C ha^{-1} yr, giving a total emission of 16 kg C ha^{-1} yr^{-1} for the study site. These estimates are conservative compared to the suggestions made by Óskarsson et al. (2004), who assumed an emission rate of 50%. Since SOC is easily transported and given the extensive accumulation of soil, the net effect of burial and subsequent reduction in decomposition is increased SOC storage (Lal, 2003). However, this increased soil accumulation decreases the quality of the soil and is also an important source of atmospheric CO_2 enrichment (Lal, 2004; Rodríguez et al., 2004).

Conclusions

The land use that initiated disturbance of soils and vegetation in Krýsuvíkurheiði was magnified by frequent volcanism during the MWP. This disturbance was later exacerbated by the LIA, in addition to anthropogenic perturbations that increased land degradation and soil erosion to a catastrophic degree. As a consequence, large amounts of aeolian material were transported over the study area during the last millennium. The aeolian processes depleted soil quality and caused significant sediment influx in vegetated areas, including the accumulation of OC and, consequently, an increase in C sequestration in soils covered by vegetation. Additional soil and SOC were lost from heavily degraded soils or barren lands. The depleted C contributed to enrichment of the atmospheric CO_2 to the extent of 16 kg C ha^{-1} yr^{-1} over 1135 yr. Despite transport of some C to the ocean and depositional sites, erosion was a net source of atmospheric CO_2, even at a conservative emission rate of 20%. Additional studies are needed on mass balance of soil and SOC and the fate of C transport by erosion with specific regard to the fraction emitted into the atmosphere.

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