The use of magnetic susceptibility to measure long-term soil redistribution

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Abstract

Several studies have documented the severity of recent soil erosion on the Canadian prairies where cultivation started about a century ago. Little quantitative information is available on erosion before 1960. This study attempts to quantify post- and pre-1960 soil erosion in a small cultivated basin near Saskatoon, Saskatchewan, Canada, by measuring \(^{137}\)Cs and magnetic susceptibility (\(\chi\)) distribution with depth. Soil cores were collected along six transects (three across closed depressions and three across the drainage channel) in the cultivated field, and one transect across an uncultivated depression. The cores were sliced into 3-cm layers and the soil analyzed for \(^{137}\)Cs, \(\chi\), and organic and inorganic C. High variability in \(\chi\) with depth in eroding areas (as indicated by \(^{137}\)Cs) made it impossible to use \(\chi\) to quantify past soil losses in these locations. However, these eroding upper and middle slope positions have a much higher \(\chi\) than lower slope areas where soil deposition occurs and where the variability in \(\chi\) with depth could be used to estimate soil deposition. Estimating soil deposition from the \(\chi\) vs. depth profiles was more successful in the closed depressions than in the drainage channel, where the \(\chi\) profiles may reflect the variable source areas of the materials rather than the pedological conditions. The data indicated that soil deposition since 1960 has been about 30 to 50% of that prior to 1960. This suggests that soil deposition rates, and by implication, soil erosion rates, have been relatively steady since cultivation of these soils started, although there are clear indications that the spatial pattern of deposition has varied. © 1998 Elsevier Science B.V.

Keywords: Erosion; Deposition; \(^{137}\)Cs; Magnetic susceptibility

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1. Introduction

Accelerated soil erosion has been recognized as a major cost to farmers on the Canadian Prairies (Science Council of Canada, 1986), but detailed field-scale information on its extent is generally lacking. Pennock and de Jong (1990), Sutherland and de Jong (1990) and others have used $^{137}$Cs to document soil redistribution since the early 1960s, but no convenient method is available to quantify the erosion that occurred during the earlier cultivation history of these soils. Magnetic techniques provide a rapid, sensitive way to characterize sediments in aquatic environments (Dearing et al., 1985; Foster and Walling, 1994), and have also been used to study soil erosion (Dearing et al., 1986; Brown, 1988). The ability of magnetic techniques to distinguish between topsoil, subsoil and parent material makes them potentially valuable for soil redistribution studies on time scales from months to millennia (Boardman et al., 1990).

The most commonly and easily measured magnetic property of soils is the magnetic susceptibility, which is defined as the ratio of the magnetisation induced in a sample to the magnetic field inducing it (Mullins, 1977): $K = M/H$, where $K$ = volume magnetic susceptibility (dimensionless), $M$ = total magnetic field per unit volume, $H$ = inducing magnetic field per unit volume. Soil scientists have commonly measured the mass magnetic susceptibility: $\chi = K/d$, where $\chi = $ mass magnetic susceptibility (m$^3$ kg$^{-1}$), $d = $ density of the material (kg m$^{-3}$). The preferred technique for measurement of magnetic susceptibility of soils involves placing the sample in a weak alternating magnetic field (Mullins, 1977).

The (mass) magnetic susceptibility of a soil is largely dependent on its content of magnetite and maghemite, since the magnetic susceptibility of these ferrimagnetic minerals is 1000 times greater than that of other iron oxides (Mullins, 1977). The occurrence of these minerals depends on the composition of the parent material and pedogenic processes, which can either increase or decrease magnetic susceptibility. Magnetic susceptibility is highest for igneous rocks and lowest for sedimentary and metamorphic rocks (Mullins, 1977). Transported parent materials can exhibit a very wide range of magnetic susceptibility (Thompson and Morton, 1979).

Topsoils generally have a higher magnetic susceptibility than subsoils because of the formation of secondary ferrimagnetics. The extent of magnetic susceptibility enhancement depends on the soil forming processes and on extreme events such as fire. Mullins (1977) has suggested that establishing a direct link between soil forming processes and a soil’s magnetic susceptibility may be impossible. It is, however, generally observed that soils in lower slope positions have a lower magnetic susceptibility than soils in upper slope positions (Le Borgne, 1954; Williams and Cooper, 1990). The objective of this study was to see if the differences in magnetic susceptibility between topsoil and subsoil, and between soils of different slope positions could be used to quantify soil redistribution in a cultivated field.

2. Materials and methods

The cultivated portion of the Floral basin, a small watershed approximately 4 km east of Saskatoon, Saskatchewan was selected for the trial. The topography of the basin is
shown in Fig. 1. The exact date when the soils of the basin were first cultivated is not known. However, the area was already well served with railways by 1910 and it is reasonable to assume that the field has been cultivated for at least 80 years. The parent material of the Dark Brown Chernozemic soils of the field is variable. A discontinuous veneer of loamy and silty glaciolacustrine deposits covers most of the area, with the till coming to the surface on steeper slopes and the lacustrine deposits occurring at the surface in low-lying areas. Using $^{137}$Cs, Martz and de Jong (1991) showed that about 90% of the field had suffered from erosion. Closed upland depressions and a series of interconnected depressions forming an intermittent drainage channel trap about 80% of the eroded soil; most of the remaining 20% is deposited in the uncultivated grassed area of the drainage channel outside the cultivated field (Fig. 1).

Fig. 1. Topographic map of the basin showing the location of the sampling transects.
Samples were taken at three locations across the drainage channel (transects T1, T2 and T3) and from three transects (T4, T5 and TX) across closed upland depressions (Fig. 1). Transects T1 to T5 were close to the sampling sites of Martz and de Jong (1991); transect TX runs east–west across the large shallow depression in the southeast part of the field. In addition, a transect (FN) across a nearby uncultivated depression was sampled. Sampling distances were from 5 to 10 m along the transects. A 6 cm-diameter soil core was taken at each sampling site with a metal probe lined with a clear plastic sleeve. Cores were described by horizons, and depth to carbonates, gleying, and sandy or gravelly layers were noted together with a designation of the slope position. The cores were cut into 3-cm segments (any surface residue was included into the 0–3 cm segment) which were air-dried and then weighed. After coarse grinding with a wooden rolling pin, gravel and stones were sieved out and weighed. The < 2 mm fine earth was weighed and used for further analysis. All data were expressed on the basis of the air-dry weight of the fine earth fraction.

The entire < 2 mm fraction from the 3 cm thick samples was put in small plastic beakers and counted for $^{137}$Cs using procedures similar to de Jong et al. (1982). Counting of each core started with the surface slice and was continued down-core until two successive core segments with no $^{137}$Cs were encountered.

Magnetic susceptibility of the core segments was measured with a Bartington Dual Frequency MS-2D meter at low frequency (0.47 kHz). Approximately 10 g air-dry soil was placed in 2.1 cm-diameter clear plastic vials to give a sample about 2 cm thick. Blank readings on an empty vial were taken before and after two readings on the sample. The two sample readings were averaged and corrected for thermal drift (using the blanks), sample weight and sample height. Corrections for sample height were based on a calibration curve obtained from a series of vials containing 0.5 cm (about 1.5 g) to 3.9 cm (about 16.5 g) of manganous carbonate.

Total and organic C were determined on all cores from the uncultivated depression and on half of the cores from the cultivated field using a LECO CR-12 carbon analyzer at temperatures of 1100°C and 820°C, respectively. Inorganic C was calculated by subtracting organic C from total C.

3. Results

3.1. Native, closed depression (transect FN)

The transect labelled FN in Fig. 2 shows the variations in $^{137}$Cs, $\chi$, organic and inorganic C for the eight cores spaced 5 m apart across the uncultivated depression. Cores FN-1 (lost) and FN-9 were on the upper slope positions and core FN-5 was from the center of the depression. Total $^{137}$Cs content of seven of the eight cores was quite similar, $2.0 \pm 0.2$ kBq m$^{-2}$ (mean ± S.D.), which is similar to the value 2.6 kBq m$^{-2}$ reported for uneroded sites by Martz and de Jong (1991), if one assumes negligible $^{137}$Cs deposition since the early 1980s. Core FN-3 contained nearly twice as much $^{137}$Cs as the other seven cores, and was also the core in which $^{137}$Cs was present to the greatest depth (10 cm). The largest $^{137}$Cs concentration was usually found in the 0–3 cm depth (Fig. 2).
but in core FN-3 it occurs at 3–6 cm depth, suggesting that this site may have received some redeposited soil. Based on the seven similar cores and assuming that up to 5% of the $^{137}$Cs falling on cultivated land could have been removed with the crop or blowing snow (de Jong et al., 1982), a $^{137}$Cs content of 1.9 kBq m$^{-2}$ would separate sites suffering a net soil loss from sites gaining soil. Since the counting errors are in the order of 10%, the following approximate soil redistribution classes were defined for the cultivated field:

- $< 1.7$ kBq m$^{-2}$: significant erosion
- $1.7$–$2.1$ kBq m$^{-2}$: little erosion or deposition
- $> 2.1$ kBq m$^{-2}$: significant deposition

The distribution of organic and inorganic C in core FN-9 (Fig. 2) is typical for thin upland Chernozemic soils. Cores FN-2 and FN-8 show increased leaching of carbonates compared to FN-9 and lack distinct Cca horizons; FN-8 is a typical orthic Chernozem. In cores FN-3 to FN-7, the Ck horizon starts at 70 cm depth or deeper, and field observations indicated gleying in several of the cores. Core FN-3 is strongly leached and has more organic C than the other lower slope cores, supporting the $^{137}$Cs evidence that it may include some redeposited soil.

The eight FN cores (Fig. 2) fall into two groups: (1) profiles with little to moderate leaching (FN-2, FN-8 and FN-9), and (2) the strongly leached lower slope profiles (FN-3 to FN-7). In the thin to moderately well-developed Chernozemic profiles of the first group, $\chi$ is generally high and seems to vary with the C content of the soil. The magnetic susceptibility of the mineral fraction of the soil is ‘diluted’ by water, humus and carbonates, all of which have a $\chi$ of $1 \times 10^{-8}$ m$^{3}$ kg$^{-1}$ or less (Mullins, 1977), in the air-dry sample. For example, in core FN-9 the Ck horizon (30 + cm depth) has a $\chi$ of about $40 \times 10^{-8}$ m$^{3}$ kg$^{-1}$ and its inorganic C content is equivalent to about 200 g CaCO$_3$ kg$^{-1}$. In the Cca horizon $\chi$ decreases to about $30 \times 10^{-8}$ m$^{3}$ kg$^{-1}$ as the carbonate content increases to about 400 g kg$^{-1}$. As CaCO$_3$ decreases to about 100 g kg$^{-1}$ in the Ah horizon, $\chi$ increases to about $65 \times 10^{-8}$ m$^{3}$ kg$^{-1}$, but it decreases in the 0–3 cm layer as the organic matter content increases to 70 g kg$^{-1}$. Thus, the enhanced magnetic susceptibility in the Ah horizon of this core seems to result at least in part from the loss of carbonates. As sand has $\chi$ values of $\sim 1 \times 10^{-8}$ m$^{3}$ kg$^{-1}$ compared with up to $15 \times 10^{-8}$ m$^{3}$ kg$^{-1}$ for common clay minerals (Mullins, 1977), the increase in sand content to 75–80% at cm depth could partly explain the small decrease in $\chi$ around that depth.

Magnetic susceptibility of the lower slope profiles (cores FN-3 to FN-7) is much lower than that of the group 1 cores, and it increases with depth (Fig. 2). This is in agreement with several studies (Mullins, 1977; Dearing et al., 1995) which have indicated that acidification, chelation and gleying appear to destroy inherited ferrimagnetic minerals or limit their pedogenetic production. Gleying was noticed in cores FN-4 to FN-6, but not for FN-3 and FN-7. The inorganic C data (Fig. 2) indicates that these two cores were strongly leached and the organic C shows some evidence of illuviation at depth.

The high variability in $\chi$ with depth in cores FN-2, FN-8 and FN-9 and the absence of a consistent enhancement in magnetic susceptibility of the topsoil, suggest that erosion cannot be estimated from changes in $\chi$. For example, erosion (and consequent
Fig. 2. Variation in magnetic susceptibility ($\chi$), $^{13}C$, and inorganic and organic C with depth along selected transects.
incorporation of subsoil by tillage) would have little effect on the $\chi$ of the A horizon of core FN-2, would increase $\chi$ in FN-8 and would decrease $\chi$ of the A horizon of FN-9. However, the much lower $\chi$ in shallower slope positions (25% or less of those of upper slopes) should allow the quantification of deposition of eroded topsoil from upper and mid slope positions. The estimation of the amount of topsoil deposited on these lower slope positions rests on two assumptions: (1) cultivation does not change the magnetic susceptibility of the lower slope soils below the depth of cultivation, and (2) the deposited topsoil does not lose its high magnetic susceptibility.

3.2. Cultivated, closed depressions (transects T4, T5 and TX)

Transect T4 (Fig. 2) consisted of eight cores spaced 10 m apart. Cores T4-1 and T4-8 are described as upper slope, cores T4-2 and T4-7 are from mid slope positions, and cores T4-3 to T4-6 are from lower slope positions. Core T4-1 is a typical Rego Chernozemic profile, lacking a B horizon and is the most eroded profile ($Cs = 1.0 \text{ kBq m}^{-2}$) of the transect. All four upper and midslope cores are in the eroded category ($Cs < 1.7 \text{ kBq m}^{-2}$) and have no $^{137}\text{Cs}$ at depths greater than 10 cm. Despite being classified as eroded, the two midslope cores (T4-2 and T4-7) have A horizons that are about 20 cm thick and B horizons extending to about 60 cm depth. Variations in $\chi$ with depth of the upper and midslope cores are in part related to inorganic C content.

Cores T4-3 to T4-6 are sites of soil deposition ($Cs > 2.1 \text{ kBq m}^{-2}$) with A horizons 18 cm or more thick. Very few carbonates were present in these four cores and there was evidence of gleying. The distribution of $\chi$ with depth is markedly different from that of the uncultivated lower slopes (cf. transect T4 vs. FN in Fig. 2). The greater $\chi$ near the surface is believed to represent deposition of topsoil from upper and midslope areas since the start of cultivation. In all four cores, $\chi$ decreases from the base towards the surface, sometimes reaching a constant minimum value in the middle of the core, and then increases to the surface. The depth at which $\chi$ starts to increase towards the surface is taken as the original depth of cultivation, and the depth to which $^{137}\text{Cs}$ is found represents the cultivation depth in the early 1960s. The current depth of cultivation is about 9 cm as indicated by the depth to which $^{137}\text{Cs}$ is found in the four eroding upper and midslope cores.

Transect TX (Fig. 2) consisted of eight cores, sampled at 10 m intervals, across a very shallow depression (Fig. 1). The cores in the central part of the depression are essentially free of carbonates and several had Ae horizons. The $^{137}\text{Cs}$ data indicated that since 1960 five (TX-1, TX-2, TX-4, TX-6 and TX-7) of the eight cores had suffered significant erosion, two (TX-5 and TX-8) showed little erosion or deposition and TX-3 showed significant deposition. The inflection point in the $\chi$ vs. depth curves support deposition of up to 17 cm of soil prior to the 1960s in core TX-3. The magnetic susceptibility data indicated 6 to 10 cm of soil was deposited at TX-4 and TX-6 before the 1960s even though these sites have eroded thereafter. Pennock et al. (1994) also reported that footslopes acted as sediment deposition areas in the first 20 years after cultivation but later suffer soil loss. Cores TX-7 and TX-8 had Ae horizons, which are associated with $\chi$ values much less than expected for middle and upper slopes. Such upland eluviated profiles are associated with minor concavities in the surface of landscapes (King et al., 1983; Miller et al., 1985).
Transect 5 started on the upper slope and ended in the center of the depression (details not shown). The upper slope core T5-1 had an Ae, but T5-2 was a more typical cultivated thin upper slope orthic Chernozem. Lower slope cores T5-3 to T5-5 all showed signs of gleying and had low values of $\chi$ in the subsoil; all three cores showed significant deposition since the 1960s (Cs > 2.1 kBq m$^{-2}$) and before.

For the cultivated depressions, Fig. 3 shows the depth to which $^{137}$Cs is incorporated and the depth of the inflection point in $\chi$ as a function of total $^{137}$Cs in the core. The depth of $^{137}$Cs incorporation in the eroding sites ($< 1.7$ kBq $^{137}$Cs m$^{-2}$) ranges from 7.5 to 11 cm, with a mean of 9 (±1) cm. In six of these cores the $\chi$ vs. depth profiles suggest that deposition had occurred prior to 1960; most of these were eluviated upland profiles. It is likely that minor surface concavities were filled in during the early years of cultivation, after which they would have become similar in erosional behavior to the surrounding upper slope profiles. Two cores showed little loss or gain of $^{137}$Cs (1.7 to 2.1 kBq m$^{-2}$); their depth of $^{137}$Cs incorporation indicates little if any soil deposition since the 1960s, but their $\chi$ profiles suggested some soil deposition before that. One of these cores is from the center of the shallow depression and the second is an eluviated upland profile. The eight cores clearly containing redeposited soil Cs ($< 2.1$ kBq m$^{-2}$) had an average depth of $^{137}$Cs incorporation of 15 cm, indicating an average of 6 cm deposition since the early 1960s. The distance between the depth of $^{137}$Cs incorporation and the inflection in the $\chi$ profile was 16 ± 8 cm, suggesting that deposition over the last three decades was about 30% of deposition over the previous five decades. Thus, there is no evidence of accelerated erosion in the last three decades compared to the immediate post-breaking period.

3.3. Transects across the drainage channel (transects T1, T2 and T3)

The main drainage channel was sampled at three locations (Fig. 1). Transect 1, which was closest to where the channel leaves the field, consisted of eight cores spaced 10 m
apart (Fig. 2). Cesium data indicated severe to slight erosion (0.7 and 1.5 kBq 137Cs m$^{-2}$) on the upper slopes (T1-8 and T1-10, respectively) to significant deposition (Cs up to 2.5 kBq m$^{-2}$) in the lower part of the channel. However, at least one of the lower slope cores (T1-6) was eroded. Martz and de Jong (1991) observed a similar pattern of adjacent erosion and deposition in the drainage channel in this vicinity.

Magnetic susceptibility of the upper and midslope cores is at least in part controlled by the amount of carbonates in the sample (cf., Fig. 2, T1-1 topsoil and subsoil and T1-3 and T1-7 variation in $\chi$ below 30 cm depth). The magnetic susceptibility in the subsoil of the three lower slope cores (T1-4 to T1-6) varies considerably, and only in T1-6 can the original depth of cultivation be established with confidence. The organic C data suggest that the original depth to cultivation was about 30 and 35 cm below the current surfaces of T1-4 and T1-5, respectively. In T1-5 this depth coincides with the inflection in the $\chi$ profile, but the major $\chi$ inflection in T1-4 is at about 80 cm depth. The soil color suggested the presence of an Ae horizon at 35 to 60 cm depth in T1-4, but the $\chi$ had not decreased to the low values associated with such horizons in the transects across closed depressions.

The 12 cores of transect 2 (details not shown) were taken 10 m apart on the upper slopes (cores T2-1 to T2-4 and T2-10 to T2-12) and 5 m apart on mid and lower slopes (cores T2-4 to T2-10). The upper slope cores ranged from severely eroded to slightly depositional. The three midslope cores (T2-4, T2-5 and T2-10) ranged from eroded to depositional. Core T2-5 contained 1.7 kBq 137Cs m$^{-2}$, but all was in the upper 3 cm, which suggests that this core has been rarely if ever cultivated; this was confirmed by the $\chi$ profile, which was very similar to lower slope cores from the uncultivated depression (transect FN in Fig. 2). The cores from the bottom of the channel all showed 137Cs accumulation and the low subsoil values of $\chi$ associated with leaching (as indicated by the absence of lime) and gleying. Except for T2-6, the original depth of cultivation could be easily estimated from the $\chi$ profiles. Core T2-10 had a $\chi$ profile with two inflection points; the upper inflection point at 23 cm was taken as the original depth of cultivation, partly based on the similarity in the $\chi$ profile with that of core T1-4 (Fig. 2).

Transect 3 (details not shown) is in the upper part of the cultivated drainage channel and consisted of eight cores spaced 10 m apart. The upper and midslope cores were moderately to non-eroded, and the lower slope cores ranged from slightly eroded (T3-5: 1.5 kBq 137Cs m$^{-2}$) to significant soil gain (T3-4 and T3-6: 3.2 kBq 137Cs m$^{-2}$). The $\chi$ profiles for the lower slope were confusing and only in T3-5 was the inflection point relatively clear. The organic C distribution suggests that originally T3-3 was cultivated to about 20 cm below the current surface, but this was difficult to infer from the $\chi$ profile.

The depth to which 137Cs was detected in the cores taken across the main drainage channel was closely related to total 137Cs content (Fig. 4) and followed the same trend as for the cultivated depressions (Fig. 3). The depth of 137Cs incorporation on the eroded sites averaged 8 cm, and was somewhat more variable (S.D. = 2 cm) than for the upland sites around the cultivated depressions. It is possible that tillage depth across the main channel is more variable than across the closed depressions. As in the closed depressions (Fig. 3), some cores with less than 2.1 kBq 137Cs m$^{-2}$ did show soil deposition prior to
1960, and for cores with more than 2.1 kBq $^{137}$Cs m$^{-2}$ there was no consistent relationship between deposition before and since 1960. For cores with more than 2.1 kBq $^{137}$Cs m$^{-2}$, deposition since the early 1960s was $9 \pm 6$ cm (using a cultivation depth of 8 cm) and deposition prior to 1960, $20 \pm 10$ cm. Thus, there is no indication that deposition rates have increased with time, implying that erosion rates have been relatively steady since cultivation started. Several of the $\chi$ profiles in the transects through the main drainage channel (e.g., T1-4 and T1-5 in Fig. 2) had complex $\chi$ vs. depth profiles, which may reflect cycles of erosion and deposition over the 10,000 years since the retreat of the last ice age, rather than pedological conditions.

4. Discussion

Figs. 3 and 4 suggest a close relationship between total amount of $^{137}$Cs in the soil and depth to which $^{137}$Cs is detected, but, the relationship between total $^{137}$Cs and depth of the original cultivation is much weaker. The magnetic susceptibility data indicated that all sites that had gained soil since the 1960s ($^{137}$Cs $> 2.1$ kBq m$^{-2}$) had also gained soil between the initial cultivation and 1960. The amount of soil deposition between 1960 and 1990 is on average 30 to 50% of that prior to 1960. Assuming that the soils were first cultivated around 1910, this suggests that the rate of deposition has not increased with time since cultivation and, therefore, that erosion rates have been steady. Although there is no indication that average deposition rates have changed markedly since cultivation started, there appear to have been shifts in the rate of deposition at various locations. Several upland cores showed evidence of soil deposition prior to the 1960s and erosion since then, and in the depositional cores there was no consistent relationship between the depths of soil deposited before and after the 1960s.
5. Conclusions

The magnetic susceptibility of uncultivated well-drained upper slope soils was several times greater than that of lower slope depressional soils. This contrast provided a means of estimating total soil deposition in lower slope positions. Deposition since the early 1960s, as estimated from depth to which $^{137}$Cs was present, was 30 to 50% of that since the soils in this area were first cultivated around the turn of this century. This suggests that soil redeposition in lower slope positions has not changed markedly since cultivation started, and implies the same for erosion on the upper slopes. The data also suggested that the spatial pattern of deposition may have changed with time.

The small magnetic susceptibilities in the lower slope soil profiles coincided with indications of high leaching (carbonates are nearly absent and often there are signs of eluviation) and gleying. The magnetic susceptibility of the majority of the topsoils in upper and midslope profiles was higher than that of the subsoils largely due to the loss of carbonates. The inconsistent enhancement of magnetic susceptibility of the upper slope topsoils coupled with variability in parent materials made it impossible to use the magnetic susceptibility data to directly estimate erosion from changes in the $\chi$ vs. depth profile.

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