



Sediment-Tracing Technology: An Overview

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PURPOSE: The purpose of this System-Wide Water Resources Program (SWWRP) technical note is to review recent literature, current trends, and research in sediment-tracing technology.

BACKGROUND: Effective watershed management can be implemented only after a thorough understanding of sediment processes in river basins. Sediment processes involve erosion, deposition, and sediment transport and, because of variability within a single watershed, are difficult to quantify.

Environmental and economic problems in water bodies can result from the lack of understanding of these processes. For example, sediment moving through a watershed transports nutrients and pollutants that can damage aquatic habitat and reduce soil productivity in agricultural areas. In addition, sediment movement and deposition in streams can increase the cost to maintain navigation and interfere with commerce and recreation.

Sediment-tracing technology began in the early 1960s with research on sediment patterns, which include erosional and depositional rates and mass sediment accumulation. The developing field of sediment-tracing technology provides a method to measure sediment erosion, estimate sediment ages, and calculate sediment deposition and accumulation rates in marine, fluvial, riverine, and lacustrine environments.

The focus of this technical note is to discuss research contributions in two areas of sediment tracing: environmental radionuclides, specifically ^{210}Pb (lead-210) and ^{137}Cs (cesium-137), and environmental magnetism. The limitations of sediment tracers and their application in different depositional environments are also addressed.

ENVIRONMENTAL RADIONUCLIDES: Natural and artificial radionuclides, collectively referred to as environmental radionuclides, are those radionuclides that are commonly occurring and widely distributed in the environment or on the landscape and are measurable (Walling 2003). Environmental radionuclides are used extensively to measure sediment movement and accumulation in closed systems or in shallow quiescent environments, such as lakes, estuaries, shallow marine, and river floodplains.

Two radionuclides, ^{210}Pb and ^{137}Cs , are well documented as producing reliable estimates of sediment accumulation rates and, hence, sediment ages. Lead-210 is a naturally occurring radionuclide, whereas ^{137}Cs is a product of nuclear testing and, therefore, an artificial radionuclide.

Lead-210 serves as a tool in reconstructing geochronology by estimating age and/or average historical sedimentation rates in sediments deposited during the last 150 years. Because of variability in a sediment column, an independent tracer, such as ^{137}Cs , is usually used in conjunction with ^{210}Pb to validate calculations derived from ^{210}Pb activity.

Lead-210 (^{210}Pb). Because of its relatively short half-life of 22.2 years (He and Walling 1996), sorbing properties to fine-grained sediment and organic matter, and measurable activity of daughter products of the ^{210}Pb radioisotope, the ^{210}Pb radioisotope proves a viable option to estimate sediment ages and sediment accumulation. Research to develop methods using the ^{210}Pb radioisotope to understand sediment processes and application of these methods over various depositional environments spans more than 40 years (Goldberg 1963; Davis et al. 1984; Appleby 1997; Appleby et al. 2003). Goldberg (1963) sparked the interest in ^{210}Pb as a viable tool to estimate geochronology by using it to determine the rate of snow accumulation in Greenland. By understanding the capabilities of ^{210}Pb to estimate geochronology, Krishnaswami et al. (1971) continued the groundbreaking research by applying the ^{210}Pb sediment age dating methods to a lacustrine environment. Models were developed and refined to accommodate different environmental and anthropogenic influences as well as geological and sediment processes, all of which contributed to variations in the sediment regime and, hence, geochronology (Pennington et al. 1976; Appleby and Oldfield 1978; Appleby and Oldfield 1983; Binford 1990; Carroll and Abraham 1996; Appleby 1997; Appleby et al. 2003; Abril 2003; Abril 2004). Research expanded to include depositional environments other than lacustrine or glacial environments (Nittrouer et al. 1979; Abril et al. 1992; Yamamoto et al. 1998; Siggers et al. 1999; Noller 2000; Panayotou 2002; Corcoran 2006). The cumulative efforts of this research represent the foundation to understand and quantify sediment processes and geochronology through the activity of the ^{210}Pb radioisotope.

The total activity of ^{210}Pb is separated into two components: unsupported ^{210}Pb activity and supported or background ^{210}Pb activity. The unsupported component is the only ^{210}Pb activity used in sediment age dating as it is considered to decay exponentially with time in accordance with its half-life (Appleby and Oldfield 1983).

Specifically, in the ^{238}U decay series, the radium isotope ^{226}Ra , with a half-life of 1,622 years, decays to the inert gas ^{222}Rn (radon), which then decays through four short-lived isotopes, each measured in a half-life of minutes, to the ^{210}Pb radioisotope (Appleby and Oldfield 1983). The decay of radon, with a half-life of 3.83 days (Gale 1995) to ^{210}Pb , begins as radon diffuses from the earth's crust into the atmosphere to produce the unsupported component of ^{210}Pb activity. As radon is isolated from its precursor by changing phases from solid to gas, the initial secular equilibrium is destroyed. A new secular equilibrium is reached as ^{210}Pb becomes the predominant radionuclide. The ^{210}Pb radioisotope then attaches to aerosols in the atmosphere to become airborne and is deposited on the earth's surface by dry fallout or precipitation. The existence of ^{210}Pb in the atmosphere is short-lived, with a mean atmospheric residence time of the order of 5 to 10 days (Krishnaswami et al. 1978). Therefore, deposition takes place on a time scale much shorter than the half-life of the isotope (Gale 1995). Approximately 90 percent of all ^{210}Pb fallout is delivered by wet deposition and, therefore, the fallout corresponds to precipitation (Binford 1993). In addition to atmospheric fallout, ^{210}Pb enters a depositional environment by runoff with subsequent deposition of suspended sediment. The ^{210}Pb produced in this process contributes to the supported component of the total ^{210}Pb activity. The unsupported ^{210}Pb

corresponds to atmospheric fallout of 0.2 to 1.0 pCi (picocurie¹) cm⁻² year⁻¹ in regions dominated by continental air masses (Binford 1993). However, the contribution of ²¹⁰Pb to the unsupported component of total ²¹⁰Pb activity declines with increasing latitude due to decreasing precipitation rates (Outridge et al. 2002).

The total inventory of ²¹⁰Pb is reported in laboratory analysis. As described by Bierman et al. (1998), ²¹⁰Pb activity is plotted directly from a laboratory analysis to produce a vertical profile, which is used to visually identify the supported ²¹⁰Pb, region of radioactivity or the unsupported ²¹⁰Pb component, and the region of sediment mixing. The difficulty in interpreting the ²¹⁰Pb activity is in isolating and identifying the components of the total inventory. The unsupported ²¹⁰Pb must be separated from the total ²¹⁰Pb, such that:

$$^{210}\text{Pb}_{\text{UNSUPPORTED}} = ^{210}\text{Pb}_{\text{TOTAL}} - ^{210}\text{Pb}_{\text{SUPPORTED}}$$

The exponential decay of unsupported ²¹⁰Pb is determined by the decay constant of the ²¹⁰Pb nuclide and thereby provides a useful dating method if the initial activity is known. Thus, the age or accumulation rate of a deposit can be determined by a general decay equation of the form (Ivanovich et al. 1992):

$$C = C_o e^{-\lambda t} \tag{1}$$

where

C_o = initial ²¹⁰Pb activity

$\lambda = \ln \frac{2}{22.26} \text{ y} = 0.03114 \text{ y}^{-1}$, the radioactive decay constant

Considerations in Application of ²¹⁰Pb. Factors influencing the accuracy of ²¹⁰Pb analysis include possible historical changes in sedimentation rate, the degree of compaction of the sediment, and the extent of post-depositional mobility, if any, of ²¹⁰Pb in the sediment column through mixing or redistribution due to biological activity, diffusion, or slumping (Farmer 1978). Depositional environments, physiographic regions, and sediment and biologic processes influence the ability of ²¹⁰Pb activity to reach substantial detectable amounts.

Sediment deposition rates in most environments are not constant so the relationship between depth and ²¹⁰Pb in a single core cannot be used to calculate apparent sedimentation rates and, therefore, a ²¹⁰Pb analysis must consist of more than one core per study area. In addition to multiple cores, certain natural conditions and sample handling methods should be considered before using ²¹⁰Pb in the calculation of sediment accumulation rates. Cores should be as undisturbed as possible and really should exhibit a uniform texture and moisture content.

¹ The following information is from the U.S. Army Corps of Engineers (2000): The curie is a standard measure for the intensity of radioactivity contained in a sample of radioactive material and the basis for the curie is the radioactivity of one gram of radium. Radium decays at a rate of approximately 2.2 trillion dpm and the picocurie is one trillionth of a curie. Therefore, a picocurie represents 2.2 dpm.

The interval in which samples are extracted from the cores is critical in defining the unsupported ^{210}Pb component, and, hence, the sediment age. Ducat and Kuehl (1995) observed that downcore ^{210}Pb cyclicity is affected by the sampling interval on cores used and noted that samples collected at 2-cm intervals throughout the length of cores give optimal representation of the nature and frequency of the ^{210}Pb cyclicity in some regions.

In an ideal situation, a uniform texture (i.e., the same grain size throughout the core) would yield an unsupported ^{210}Pb activity that is readily identified. This geologic situation is unlikely in most environments. Although research by Kuehl et al. (1986) and Underkoffler (1990) found that a relationship does exist between excess ^{210}Pb activity and mean grain-size, Ducat and Kuehl (1995) disputed that such a relationship exists. Further research is needed in this area to qualify the mineralogical and environmental influences on ^{210}Pb activity.

As with all data based on cores, whether disturbed or undisturbed, sediment compaction presents a problem when correlating ^{210}Pb activity with associated sediment ages (Martin and Rice 1981). Farmer (1978) studied the effects of sediment compaction on sedimentation rates and applied a depth correction by normalizing wet-dry ratios (on a volume basis) in uncompacted sediments to that of fairly constant wet-dry ratios at depth.

The initial concentration of ^{210}Pb in the sediment is low when the accumulation of sediment is rapid and is high when it is slow (Goldberg 1963). Usually, an accumulation rate of greater than 1 mm/year is considered sufficient for ^{210}Pb dating (Oldfield and Appleby 1984). In areas where sediment accumulation is slow and the degree of reworking of the sediment column is intense, accurate measurement of accumulation rates may not be possible (Peng and Broecker 1979). Bioturbation, erosion, and mixing by wave or current action are post-depositional processes that disturb the sediment and interfere with the accuracy of ^{210}Pb sediment age dating. Cores with complex stratigraphy, such as those with interbedded sands and muds, can be of limited use when calculating sedimentation rates due to mixing of fine- and coarse-grained soil. In addition to these disturbances, Binford (1990) addressed natural variation, the statistical nature of measuring radioactivity, and the estimation of supported ^{210}Pb that influence ^{210}Pb sediment age dating as sources of error.

Research concludes that the maximum usefulness of ^{210}Pb in estimating sediment processes is in quiet, unmixed water deposits of continuous sedimentation over the past 150 years (Appleby 1997; Noller 2000; Corcoran 2006). In addition to identifying ideal depositional and environmental targets, other research highlighted geographic and vegetative influences. Branford et al. (1998) found that complex topography correlates to increased ^{210}Pb activity. In areas with a dense forest canopy, especially conifer forests, the level of ^{210}Pb activity is higher than that found in soils with minor or no canopy (Yamamoto et al. 1998).

Description and Application of ^{210}Pb Models. The total ^{210}Pb activity follows an exponential decay with depth, assuming a uniform sedimentation rate over a 150-year period. However, because of anthropogenic and environmental effects, this is seldom the case. Several models for ^{210}Pb analysis have been developed using the radioactive decay equation. Model selection depends on environmental conditions, sediment processes (e.g., erosion, deposition), sediment focusing, and sediment stability. To select a model for ^{210}Pb sediment chronology, a historic knowledge of the sampled sediment is needed. The limitations of these models are due

primarily to selecting the background level of ^{210}Pb and the number of samples analyzed. The age, t , is calculated if the initial ^{210}Pb activity can be derived from a suitable model (Appleby and Oldfield 1992). Based on assumptions inherent with any sediment dating models, the derived dates can only be considered as estimates.

As with all models, there are certain assumptions that are accepted, and specific models may have additional assumptions and constraints. Pennington et al. (1976) and Noller (2000) note the general assumptions as follows:

- ^{210}Pb is quickly removed from the atmosphere and freshwater streams and sequestered in soils and sediments.
- ^{210}Pb is immobile once deposited.
- Unsupported ^{210}Pb does not migrate down into the sedimentary column.
- Supported ^{210}Pb is in secular equilibrium with its grandparent ^{226}Ra .
- Unsupported ^{210}Pb is independent of depth.

In accordance with the radioactive decay equation, the exponential decay of ^{210}Pb should display a linear, monotonic profile (Appleby and Oldfield 1983; Abril 2004). However, researchers have observed nonmonotonic and nonlinear ^{210}Pb activity profiles produced from depositional and erosional cycles (Appleby and Oldfield 1983). The problem arises when the same ^{210}Pb data produce different results calculated from different models (Abril 2004). Because of this, it is necessary to validate the ^{210}Pb geochronology with a time-dependent marker. An event tracer, such as ^{137}Cs , should be plotted with ^{210}Pb model dates to produce a more effective model of age estimation.

The Constant Flux:Constant Sedimentation (CF:CS), Constant Rate of Supply (CRS), and Constant Initial Concentration (CIC) models are used to calculate sediment age at sample depth. The main distinction between the latter two is that the CRS assumes a constant atmospheric flux of ^{210}Pb and the CIC is governed by a varying sedimentation accumulation rate that does not affect the ^{210}Pb concentration. If a ^{210}Pb activity-to-depth profile is nonlinear and strongly non-monotonic, then only the CRS model offers a method to interpret the stratigraphic chronology. Although corrected CRS model dates can be calculated ^{137}Cs , errors can still arise due to variations in the ^{210}Pb supply rate at individual core sites from local irregularities in the sedimentation process (Appleby 2000). The CF:CS, CRS, and CIC are introduced at the inception of ^{210}Pb model development and, therefore, applied more often than other models.

The ^{210}Pb models are described in the following sections.

Constant Flux and Constant Sedimentation rate (CF:CS) model. Mass sediment accumulation rates are a measure of sedimentation where there are changes with depth in sediment density that relate to compaction or changes in sediment composition. These sedimentation rates are determined by plotting the log of the excess ^{210}Pb activities against cumulative dry sediment weights and calculating sedimentation rates over intervals of constant slope (Brush et al. 1982). The sedimentation rate can be calculated using the slope of the line derived from the linear regression of $\ln^{210}\text{Pb}_{\text{UNSUPPORTED}}$ and sample depth in the following equation (Bierman et al. 1998):

$$A(z) = A(o)e^{-bz} \quad (2)$$

The rate of sedimentation, S , is equal to λ/b in cm/year, where

- λ = ^{210}Pb radioactive decay constant of 0.03114
- $A(z)$ = excess ^{210}Pb activity at depth z
- $A(o)$ = activity at the surface (or bottom of the uniformly mixed layer)
- b = the slope defined by a regression through the data
- z = depth

Assuming a constant activity at the surface, the model is simplified to the following equation:

$$S = \frac{\lambda_{210\text{Pb}}}{b} \quad (3)$$

where the slope (b) of the regression line through data is plotted as the natural log of unsupported ^{210}Pb activity as a function of sample depth.

Constant Initial Concentration (CIC) model. The CIC model, also known as the constant specific activity model, assumes that an increased flux of sedimentary particles from the water column will remove proportionally increased amounts of ^{210}Pb from the water to the sediments (Appleby and Oldfield 1983) and will increase or decrease in response to changes in sedimentation. The CIC model assumes that the vertical sediment accumulation rate will not affect the ^{210}Pb concentration and that it will remain constant. If the initial concentration in the sediment is low, the unsupported ^{210}Pb will require less time to decay to a level indistinguishable from the supported ^{210}Pb than if it was initially high (Pennington et al. 1976). The Constant Initial Concentration (CIC) model assumes that an increased flux of sedimentary particles from the water column will be removed proportionally. Appleby (2001) observed that discrepancies with the CIC model date indicate variations in the initial ^{210}Pb concentration and can be due to a variety of causes, such as flood events, sediment slumps, turbidity events, and major land-use changes.

The unsupported ^{210}Pb concentration will vary with depth in accordance with the formula (Appleby and Oldfield 1983):

$$C = C(o)e^{-\lambda t} \quad (4)$$

where $C(o)$ is the unsupported ^{210}Pb concentration of sediments at the sediment water interface and the radioactive decay constant (λ) of ^{210}Pb is:

$$\lambda_{210\text{Pb}} = \frac{\log 2}{22.26} = 0.03114\text{y}^{-1}$$

$$\text{or } \lambda_{210\text{Pb}} = \ln \frac{2}{22.26\text{y}} = 0.03114\text{y}^{-1}, \text{ as previously stated.} \quad (5)$$

According to Appleby and Oldfield (1983), the age of sediment layers with ^{210}Pb concentration C is therefore:

$$t = \frac{1}{\lambda_{^{210}\text{Pb}}} \ln \frac{C(o)}{C} \quad (6)$$

Constant Rate of Supply (CRS) model. Based on the law of radioactive decay, the CRS model, also referred to as the Constant Flux model, is used to determine the age of a given depth from a ^{210}Pb vertical profile within a sediment column. The CRS model is the most widely accepted model and is dominated by constant direct atmospheric fallout (Appleby et al. 1990). The model is based on assumptions that the unsupported ^{210}Pb is supplied at a constant rate to sediments through time, the initial ^{210}Pb concentration in the sediment is variable, and the influx rate of sediment is variable (Goldberg 1963; Appleby and Oldfield 1978; Noller 2000). Because the CRS model assumes constant fallout of ^{210}Pb , it also considers that no post-depositional mixing occurs. Another assumption is that the transport parameters are independent of sedimentation rate. Given these inherent assumptions, the CRS model is usually appropriate in bodies of water where there has been little or no sediment disturbance. Sediment mixing typically results in a flattening of the profile of ^{210}Pb activity versus depth in the surficial sediment layers, and degradation of the ^{137}Cs peak (Appleby 2001). In the CRS model, the sediment accumulation rate can vary.

According to Appleby and Oldfield (1978), the initial concentration $C_o(t)$ of unsupported ^{210}Pb in sediment of age t years must satisfy the following:

$$C_o(t) r(t) = \text{constant} \quad (7)$$

where $r(t)$ ($\text{g}/\text{cm}^2\text{yr}$) is the dry mass sedimentation rate at time t .

From this equation, Appleby and Oldfield (1978) developed a relation for the age of deposit at depth x :

$$t = \frac{1}{\lambda} \ln \left(\frac{A_0}{A_x} \right) \quad (8)$$

where A_0 is the total unsupported ^{210}Pb activity in the sediment column and A_x is the total unsupported ^{210}Pb activity in the sediment column beneath depth x (Noller 2000).

The sedimentation rate, r , is calculated directly from the formula (Appleby and Oldfield 1978):

$$r = \frac{\lambda A_x}{C} \quad (9)$$

The net flux of ^{210}Pb to sediments is calculated by the following equation (Binford 1993):

$$P = \sum ^{210}\text{Pb}_{\text{UNSUPPORTED}} \lambda \quad (10)$$

where $\Sigma^{210}\text{Pb}_{\text{UNSUPPORTED}}$ is the sum of the unsupported activity and P is the net flux. The result should not exceed the global ^{210}Pb natural atmospheric flux of 1.3 to 5.8 pCi in⁻² y⁻¹ (Appleby and Oldfield 1983) providing an indication of the validity of ^{210}Pb observations.

Cesium-137 (^{137}Cs). In 1945, the United States began atmospheric nuclear testing (Durham and Joshi 1980; Kathren 1984). Fallout radionuclides, including ^{137}Cs , were released into the atmosphere, deposited on the earth, and quickly adsorbed onto sediments. The testing intensified in 1958 and 1963, and these peaks of ^{137}Cs activity can be used to measure and date sediment; measure and predict erosion; and using these data, predict the origin of sediment (Mitchell et al. 1983) provided no post-depositional mixing has occurred. These peaks in the sediment record are also used to calibrate other sediment tracer methods, such as the ^{210}Pb radioisotope.

Research to define the relationship between fallout radionuclides and soil erosion and transport began in the early 1960s (Ritchie and McHenry 1990). In the early 1970s, extensive research was conducted on the use of ^{137}Cs in dating sediment. McHenry and Ritchie (1977) used the spatial distribution of ^{137}Cs to show erosion patterns in fields, redistribution of soils within fields, and soil loss from fields. The time distribution of radioactive fallout of ^{137}Cs has been measured extensively and documented for locations globally (Cambray et al. 1983).

Appleby and Smith (1993) described the value of comparing ^{137}Cs and ^{210}Pb ; first, because the response of ^{210}Pb to transport processes may be different from that of ^{137}Cs , comparisons between the two will be helpful in validating the model and, secondly, because radionuclide inventories may be affected by sediment processes. Normalization of ^{137}Cs inventories against ^{210}Pb will to some extent allow an elimination of the effects of such processes.

The ^{137}Cs radioisotope, with a half-life of 30.2 years (He and Walling 1996), possesses specific properties, which contribute to using it as a viable sediment tracer. According to Ritchie and McHenry (1990), these properties are as follows:

- a. There are no natural sources of ^{137}Cs .
- b. Fallout is strongly related to local precipitation patterns with total fallout varying linearly with rainfall within latitudinal zones.
- c. Temporal and spatial distribution of radioactive fallout on the earth's surface has been extensively measured and documented.
- d. ^{137}Cs is strongly adsorbed on clay and organic particles and is essentially nonexchangeable.
- e. Adsorption on soils and sediments is rapid with distribution in undisturbed soil profiles showing an exponential decrease with soil depth.

The use of ^{137}Cs in erosion studies depends on measuring the decline of ^{137}Cs fallout in surface soil to quantify surface erosion. Because ^{137}Cs is deposited as fallout and is strongly adsorbed to soil, its movement by chemical and biological processes can occur but is limited (Ritchie and McHenry 1990). By measuring the spatial patterns of residual ^{137}Cs across a landscape, rates of soil loss and deposition for a field or parts of the field can be estimated to provide an estimate of net soil loss at each site (Nagle and Ritchie 2004).

River floodplains represent important sediment sinks for storing suspended sediment and contaminants from upstream. The ^{137}Cs radioisotope is an established sediment tracer used to study floodplain deposits and sediment sources (Owens et al. 1999; Walling and He 1997). As with the ^{210}Pb radioisotope, the ^{137}Cs activity peaks proved useful in estimating sediment ages and deriving sedimentation rates in a lacustrine environment (Pennington 1973; Ritchie et al. 1983; Appleby et al. 2003).

Considerations in Application of ^{137}Cs . The basis for using ^{137}Cs for deriving sedimentation rates is that it is rapidly and strongly adsorbed by soil, preferentially by clay-sized particles (Tamura 1964). The ^{137}Cs presence in soils is from direct deposition from the atmosphere, from vegetation, turnover from vegetation, redistribution of eroded particles, and deposition from water on floodplains (Walling and He 1993). The presence of ^{137}Cs in water is from direct deposition on water surfaces and input of ^{137}Cs adsorbed on eroded soil particles (Ritchie and McHenry 1990). The amount of ^{137}Cs contributing to aquatic or terrestrial environments is also dependent on precipitation rates (Mitchell et al. 1983).

The ^{137}Cs fallout deposited on vegetation may be adsorbed or absorbed. Most adsorbed ^{137}Cs is washed from vegetation and moved to the soil. Absorbed ^{137}Cs is released to soils when vegetation dies and decays (Ritchie and McHenry 1990).

As with all sediment tracers, errors have been identified with ^{137}Cs sediment age dating. Davis et al. (1984) found that ^{137}Cs activity, when compared to pollen analysis in undisturbed sediment, was present at pre-fallout sediment depths. Because the sediment exhibited no post-depositional mixing by bioturbation, Davis et al. (1984) suggested that ^{137}Cs undergoes diffusion in the sediment column.

When using ^{137}Cs as a sediment tracer, Ritchie and McHenry (1990) sum up the considerations as follows:

- a. Only two dates (1954 and 1963) can be determined for most sediment in the northern hemisphere although, because of the short half-life of ^{137}Cs , the 1954 horizon is difficult to identify.
- b. ^{137}Cs is preferentially adsorbed on clay particles. In areas where sand is prevalent, there will be a decrease in the amount of ^{137}Cs present.
- c. There is a time lag between the time of atmospheric deposition of ^{137}Cs and the time of deposition in soil.
- d. Diffusional movement of ^{137}Cs can occur and may redistribute ^{137}Cs in sediment profiles.
- e. ^{137}Cs is not useful in sediments that have been disturbed by bioturbation or tillage or in areas where the sedimentation rate is <1 cm per year.

Estimation of Sediment Age from ^{137}Cs Profiles. While ^{210}Pb is used to date an entire sediment column with accumulation in the last 150 years, ^{137}Cs is an event marker, used only for specific dates of 1954, the onset of ^{137}Cs fallout, and 1963, the peak ^{137}Cs fallout. Profiles of ^{137}Cs activity are made directly from laboratory analysis. Discrepancies between ^{210}Pb and ^{137}Cs are perhaps due to the downward migration of ^{137}Cs in the sediment column. Although ^{137}Cs is

used as an independent tracer in conjunction with ^{210}Pb , errors may still arise due to variations in the ^{210}Pb supply rate from local irregularities in the sedimentation process (Appleby 2000).

ENVIRONMENTAL MAGNETISM: All substances possess magnetic properties (Walden et al. 1999). Environmental magnetism is based on the premise that natural materials behave in characteristic and unique ways in the presence of magnetic current. This observed signature behavior becomes another physical property that can be used to identify minerals in sediment tracer studies. In the words of the pioneers in the field, Thompson and Oldfield (1986), “magnetic parameters can be diagnostic of mineral type and origin; they are sensitive to chemical and thermal transformations and can reflect the ambient magnetic field at certain critical stages in processes such as sediment deposition, rock cooling and crystal growth.” When paired with radioisotope geochronology, pollen analysis and geomorphic studies, the ability to characterize sediment in dynamic environments such as lakes and streams is greatly enhanced.

The development of environmental magnetism in sediment studies began in the late 1960s at Lough Neagh, Northern Ireland (Walden et al. 1999). This research led to the interpretation of both natural remanence and laboratory-induced magnetic properties such as magnetic susceptibility and isothermal remanence (Oldfield 1991). Magnetic susceptibility was used by Thompson et al. (1975) to characterize and identify the source of lake sediments.

Many sediment-tracing studies have investigated links between sediment in lakes and the sources in the drainage basin (Thompson and Morton 1979; Walling et al. 1979; Oldfield et al. 1979; Bradshaw and Thompson 1985; Oldfield et al. 1985; and Dearing 1992).

Royall (2001) combined ^{137}Cs and mineral magnetic measurements to hillslope sediment to study sediment delivery to a small stream. Caitcheon (1998) used sediment tracing to evaluate sediment contribution at stream junctions. Dearing (2000) reviewed the current use of naturally occurring Fe-bearing minerals as chemical tracers in fluvial systems. The sediment in question must have a clear and unambiguous magnetic signature. This may represent a specific mineral. Fingerprinting is an approach to sediment sourcing that uses mineral magnetic analyses to provide a compositional fingerprint by which catchment sources can be compared with suspended sediments (Oldfield et al. 1979; Walling et al. 1979; Caitcheon 1993; Slattery et al. 1995).

Mineral magnetic properties have applications in defining the presence of contaminated sediment (Pozza et al. 2004). Limitations to this method are that the magnetic values cannot directly determine the actual pollutant concentrations and the method works only in areas where the bottom sediments have low magnetic susceptibilities.

Basic Magnetism. Dearing (1999) states that for each substance there is a relationship between the magnetic field and the amount of magnetization created. This relationship can be defined by the ratio of the two: the strength of the magnetization divided by the magnetic field. The ratio is the magnetic susceptibility and is expressed as:

$$K=M/H \tag{11}$$

where K is volume susceptibility, M is the magnetic moment per unit volume of sample, and H is the applied magnetic field (Dearing 1999). Measurements of magnetic susceptibility can be made under different conditions to further describe the type, quantity, size, and orientation of minerals in the sample.

When a magnetic field is applied to a material, certain magnetic properties can be measured. These characteristic properties have been outlined by Thompson and Oldfield (1986) as:

- a. **Diamagnetism:** When a magnetic field is applied, a weak negative magnetization occurs. It is present in many common natural materials such as quartz, feldspar, calcite, and water. When the magnetic field is removed, diamagnetism is lost.
- b. **Paramagnetism:** When a magnetic field is applied to a material, a weak positive magnetization occurs. Some thermal agitation results and will eventually break down the magnetization. It is present in olivine, pyroxene, garnet, biotite, and carbonates of iron and manganese.
- c. **Ferromagnetism:** Characterized by remanent magnetization or the ability to retain some magnetization after the removal of a magnetic field. Further characterization is done by observing the way that the magnetic properties change with temperature. At a certain critical temperature for that particular material, called the Curie temperature, the remanent magnetism is lost. Iron minerals belong to this group,
- d. **Ferrimagnetism:** Similar to ferromagnetism except that there is a net magnetization that is unique to that material. Ferrimagnets also have low electrical conductivities. Iron oxides are in this group.

Another characteristic property that is measured and described is hysteresis or the behavior of a specimen as magnetic fields are applied and then reapplied. A magnetization curve is created that is unique for the material being tested (Thompson and Oldfield 1986).

Considerations in Application of Magnetic Properties. Determining the sources of sediment in a basin can become complex, mathematically and statistically. Some magnetic properties can be lost with erosion processes. Magnetic minerals are created and can be diagenetically altered through environmental processes. The process of dissolution leads to the loss of the finest magnetic grains first, followed by the coarser grained material (Anderson and Rippey 1988; King and Channell 1991). Marine and freshwater bacteria have the ability to dissolve and reduce magnetite in the soil (Kostka and Nealson 1995; Paasche et al. 2004). Authigenic growth of magnetic minerals can also occur in both freshwater and marine environments (Hilton 1990; Roberts and Turner 1993).

Breakage and/or erosion have substantial effects on the magnetic properties of soils. Abrasion and breakage of sediment particles occur during transport and will affect the concentration of magnetic minerals. The net effect with soil from the erosion of granitic rocks is to reduce the concentration. The net effect with soil originating from sedimentary rocks is to accumulate particles with higher magnetic concentrations (Crockford and Olley 1998).

SUMMARY: The dynamic field of sediment tracers covers a broad range of artificial and natural radionuclides as well as environmental properties of sediment. The techniques described in this SWWRP technical note offer researchers an opportunity to expand their knowledge of

sediment tracers. Understanding the application of these tracers advances the understanding of sediment processes in different depositional environments.

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