

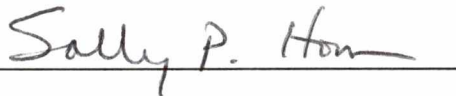
To the Graduate Council:

I am submitting herewith a dissertation written by Phillip Daniel Royall entitled "Lake-Sediment-Based Evaluation of Sediment Yield and Dynamics, Crooked Run Drainage Basin, Virginia." I have examined the final copy of this dissertation for form and content and recommend that it be accepted in partial fulfillment of the requirements for the degree of Doctor of Philosophy, with a major in Geography.

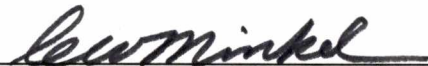


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We have read this dissertation and
recommend its acceptance:



Accepted for the Council:



Associate Vice Chancellor and
Dean of the Graduate School

LAKE-SEDIMENT-BASED EVALUATION OF SEDIMENT
YIELD AND DYNAMICS, CROOKED RUN DRAINAGE BASIN,
VIRGINIA

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ABSTRACT

Because of the potential for sediment storage in watersheds over a variety of timescales, and because the mobilization of stored sediment is related to the frequency-magnitude characteristics of runoff events, short-term sediment monitoring data cannot be assumed to apply beyond the period of measurement. Medium-term (10^1 - 10^2 years) sediment yield data are required to understand the dynamics of sediment storages as well as the persistent effects of natural and anthropogenic watershed disturbances. In this dissertation I present a 29-year sediment yield record for a small watershed in Virginia derived from analyses of reservoir sediment. The sediment yield record is evaluated with regard to precipitation and discharge data from nearest monitoring stations and to dendrohydrological evidence of past flooding within the catchment.

The sediment yield reconstruction was accomplished by correlating stratigraphic layers within a network of lake sediment cores, calculating the mineral mass of each layer, and establishing an absolute chronology for one representative master core in the network. Measurements of magnetic susceptibility, loss-on-ignition and radionuclide content of one-centimeter sediment increments were used to accomplish this reconstruction. The resulting sediment yield record isolates four broad periods of sediment yield as well as sediment contributions from two individual large storm events. Average values of sediment yield for each of the primary sediment yield periods appear related to annual precipitation characteristics at the site. I use the storm event sediment yields and the 29-year average to estimate the relative contributions of large infrequent floods and frequently occurring moderate floods to medium- and long-term sediment yield. The sediment yield contribution from the largest storm event forms only 10% of the total 29-year sediment accumulation, and much less when the recurrence interval of the associated flood event is accounted for. This importance value is similar to that published

for much larger streams in the area, and therefore does not support the theory that the importance of sediment yield contributions from high-magnitude events increases as basin size decreases.

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LIST OF SYMBOLS AND ABBREVIATIONS

LOI	Loss-On-Ignition (analysis).
L(#)	Lake bank erosion pin number.
P(#)	Channel bank erosion pin number.
phi	Standard particle size unit; = $-\log_2$ (grain diameter in mm).
USGS	United States Geological Survey.
USFS	United States Forest Service.
X_{lf}	Mass specific low frequency magnetic susceptibility ($m^3 \text{ kg}^{-1}$).
X_{fd}	Frequency dependent magnetic susceptibility (%).
Z,A,B, C,C/D, D,E,F(#)	Coring grid designations for retrieved cores.

I. INTRODUCTION

Accurate estimates of sediment yield from watersheds are important for identifying and managing potential water quality hazards, quantifying soil depletion rates on cropland and land managed for timber, understanding the effects of natural watershed disturbance, and assessing landform change (Hadley, 1980; Foster et al., 1988; Richards, 1993). Despite advances in predicting the hydrological behavior of watersheds, attempts to model temporal variation in sediment yield have met with limited success because the controlling factors remain poorly understood (Walling and Webb, 1983). The temporal dimensions of these factors are particularly difficult to understand because lengthy sediment yield records are generally unavailable for analysis (Dearing et al., 1982). Most sediment yield data have come from short-term (less than five-year) monitoring programs providing measurements of suspended sediment concentration in stream water (Dearing et al., 1982). Because of the potential for sediment storage in watersheds over timescales from months to thousands of years, and because the remobilization of stored sediment is related to the frequency and magnitude of runoff events (Meade et al., 1990), the results of short-term monitoring programs cannot be assumed to apply beyond the period of measurement. Sediment yield data covering time periods in excess of five years are necessary for understanding soil erosion and sediment storage dynamics, and the impact of anthropogenic and natural disturbances that may have persistent effects on watersheds. Dunne (1984) and Hadley (1985) emphasized the lack of medium-term (10^1 - 10^2 years) sediment yield data as an important problem for future watershed research, and have pointed to the need for longer sediment yield records from both natural areas and areas substantially altered by human activities.

Natural lakes and artificial water impoundments (reservoirs) serve as sediment traps for the watersheds that drain into them, and therefore potentially record changes in

sediment yield since their creation (Bondurant and Livesy, 1973; Oldfield, 1977; Foster et al., 1985). Methods used to measure lake sedimentation range from simple bathymetric surveying (e.g., Borland, 1971; Jolly, 1982; Garcia and Vignoli, 1988) to detailed compositional analyses of lake sediment cores (Davis, 1976; Foster et al., 1985). The application of isotopic dating techniques to recent lake sediment (Ritchie et al., 1973; Oldfield and Appleby, 1984) and advances in techniques of core correlation (Thompson and Oldfield, 1986) have made it possible not only to derive average values of sediment yield since a lake's impoundment, but also in some cases to provide estimates of changes in sediment yield through time. This approach has most often been used to link changes in sediment yield to contemporaneous changes in land use affecting lake sediment accumulation occurring over a period of several years or more (Stott, 1987; Dearing et al., 1981; Dearing, 1992; Flower et al., 1984; Dearing et al., 1987).

In contrast, Foster et al. (1988) have noted the potential of lake sediment-based techniques to advance research related to the frequency and magnitude of sediment transport events that commonly last from a few hours to a few days. The basic question is, what proportion of the total medium- and long-term catchment sediment yield is attributable to storm events of high magnitude and low frequency? Conversely, how important to the total medium- and long-term sediment yield are frequently occurring small to moderate storms? Although the potential usefulness of lake sediments for answering these questions has been recognized (Foster et al., 1988), the topic remains essentially unexplored.

On June 22, 1972, Hurricane Agnes produced record-breaking rainfall and flooding through parts of central and northern Virginia as it moved along the eastern United States. The recurrence interval of this rainfall event and of the resulting flooding are estimated at greater than 100 years in affected areas of Virginia (Bailey et al., 1975). Although the degree of landform modification by this event was smaller than expected at

some sites (Moss and Kochel, 1978), sediment concentrations in stream water were unusually high in most areas (Bailey et al., 1975). This observation, however, is biased by spatial scale because most sediment monitoring stations are located on major alluvial trunk streams. Much less is known about the response of smaller catchments to such large storms in terms of sediment yield. Knowledge of these effects is important because headwater catchment processes are significant both in terms of conditioning responses downstream and in terms of reflecting more direct linkages to upland areas affected by various disturbances within the drainage system. How much sediment yield from headwater basins was associated with Hurricane Agnes and other high magnitude storms? How do these values compare to medium-term average sediment yields from the same basins? Do these frequency-magnitude characteristics themselves vary over space? If so, what are the controlling factors? These questions require sediment yield information spanning periods longer than that encompassed by direct monitoring programs and also more widely distributed in space.

In this dissertation I present a record of sediment yield from the Thompson Lake catchment located within the area affected by Hurricane Agnes. The record spans the last 29 years of sediment yield from this small unmonitored mountain catchment, incorporating the year (1972) of Hurricane Agnes, and is derived from detailed analyses of sediment cores extracted from the lake. The general objectives of the study are to obtain information regarding the amount and cause of change in average sediment yield from the Thompson Lake watershed since its impoundment, and the frequency and magnitude characteristics of fluvial sediment export. I derive this information by delineating and quantifying the mineral mass of traceable lake sediment layers representative of different depositional periods at the site, and layers that are directly attributed to individual flood events. If individual floods can be identified in the lake sediment record, future application of the methodology to unmonitored lake catchments

elsewhere in the region should help reveal spatial controls of sediment yield frequency and magnitude. Although the 29-year age of the reservoir is small compared to that of most earlier similar studies, and is not sufficiently long to span several periods of land use change, it is long enough to register changes in sediment yield associated with intrinsic watershed dynamics over a period of only mild upland disturbance.

In addition to providing information on medium-term sediment yield averages and frequency-magnitude issues, this research makes further contributions to the field of watershed dynamics. The record from Thompson Lake represents the first reservoir-based sediment yield record from the eastern United States using sediment stratigraphy rather than whole lake volumetric resurvey assessments. It adds to the small number of headwater catchments in the southern Appalachians for which sediment yield data are available. It adds to the even smaller number of small catchments for which as many as 29 years of sediment yield are represented in the record. Finally, it is the first lake catchment-based sediment yield study that incorporates evidence of past flooding derived from tree-ring analysis to aid in the reconstruction of flood history at a site.

The significance of these contributions to watershed research is best understood by placing them within a proper disciplinary context. Accordingly, the following chapter provides background information on the general nature of watershed sediment dynamics, the influence of natural and anthropogenic disturbances on sediment yield, current research issues in sediment yield, and the particular approach to studying sediment yield adopted in this research. The succeeding two chapters address in greater detail the site environment and the research methods used to obtain watershed process information both directly from the catchment and through the study of lake sediment. The next three chapters detail the specific results of applying these methods and address the data derived from catchment mapping and monitoring, flood history reconstruction, and the lake

sediment-based sediment yield record. The final two chapters provide an evaluation and interpretation of these results and a summation of the important conclusions reached.

II. BACKGROUND

Dynamics of Sediment Yield

Annual sediment yield is defined as the mass of sediment leaving a drainage basin over one year (Richards, 1993). In humid environments, removal of material in solution may account for a large portion of total sediment yield (Velbel, 1985). With respect to land surface modification, however, solution is more important for its weathering rather than its denudational effects because, outside of carbonate or other soluble rock areas, solutional weathering and erosion tends to be isovolumetric (Velbel, 1985). The following discussion is focused on the particulate portion of stream sediment loads, although much of the information presented applies also to dissolved load.

Sediment yield is controlled by watershed variables including geology, catchment morphology, climate, soil properties, land use and other factors influencing the type and rate of hydrologic processes and the type of material available for erosion. Bedrock mineralogy and structure, topography, climate and biota interact to determine soil properties such as profile thickness, texture, structure, infiltration capacity and organic matter content that affect the erodibility of the land surface. Erosion of soil particulates may occur via moving water (stream and sheetwash entrainment, rainsplash), mass wasting (e.g., soil creep, landsliding), wind (eolian entrainment), and animal activities (e.g., animal burrowing). Frequency-magnitude characteristics, intensity and duration of precipitation events affect the location and duration of erosive overland and subsurface flow on hillslopes, the associated characteristics of flow in the channel network, and the frequency of catastrophic mass-wasting events (Dietrich and Dunne, 1978; Dietrich and Dunne, 1993). Eolian processes may remove and redistribute soil particles, but are generally of lower importance relative to hydrologic processes in forested watersheds. In

contrast, the burrowing activity of animals is capable of moving sizable amounts of topsoil downslope towards stream channels (Imeson, 1976; Thorn, 1978). The combined result of these and other weathering, erosion and deposition processes is the production of loose sediment available to be reworked in stream channels and manifest in the form of sediment yield downstream.

Sediment availability rarely translates directly into sediment yield because of the potential for temporary and long-term storage of mobilized sediment on hillslopes, on floodplains, and within channels (Walling, 1983; Richards, 1993). Barriers to predicting sediment storage dynamics include poor understanding of the mechanisms behind sediment transfer into storage, lack of knowledge regarding the location and volumes of storages, and a lack of knowledge concerning the residence time of mobilized sediment in different storage locations (Richards, 1993). The length of sediment yield record required to address the latter question must reflect the travel time of sediment through temporary active storages within a channel (Richards, 1993). The residence time constituting temporary active storage is logically defined by the timescale of interest. For questions concerning water quality or anthropogenic influences for example, an historical timescale (less than 200 years) might be appropriate. At such a scale, sediment considered to be in active storage may be remobilized by flow events with recurrence interval of less than five years (i.e., near the average annual flood recurrence interval). Unfortunately, travel time of sediment in active storage varies between catchments, although it is likely to be proportional to catchment size (Richards, 1993).

A key theoretical debate related to residence time of stored sediment involves the efficacy of flow events of variable frequency and magnitude, with respect to medium- and long-term sediment yield. The basic question is: do infrequent but high-magnitude events account for most of the earth's denudation, or does the high frequency of smaller events compensate for the smaller amount of work done per event? Wolman and Miller

(1960) theorized that most denudation is accomplished by intermediate-sized events of moderate frequency and magnitude. Their argument assumes: 1.) the rate of sediment movement is a power function of applied stress (proportional to discharge) and 2.) the frequency distribution of flood peaks of different magnitudes is near log-normal. Under these circumstances, the product of rate and frequency (i.e., the sediment yield, or work accomplished) attains a maximum value that is centered over the moderate stress (moderate magnitude and frequency) portion of the frequency distribution. This result holds only for cases in which the threshold stress for sediment movement is exceeded. Wolman and Miller (1969) provide field examples documenting the recurrence interval (the reciprocal of frequency) at which work is maximized. Two basins, one semi-arid and one humid in climate, are provided as examples. Ninety percent of sediment yield from both basins is accounted for by flows with recurrence interval of less than five years. At these sites 50% of sediment yield was associated with flows occurring on average one or more days per year. Flows with recurrence interval of less than five years were considered by Wolman and Miller (1960) to be of moderate frequency and magnitude.

This frequency/magnitude relationship is predicted to vary with basin area because the steadier the flow, the steadier the rate of sediment transport. Small headwater catchments typically experience greater variation in flow than larger river segments downstream and therefore are presumed to have a greater proportion of their total load transported in larger, less frequent events (Leopold, Wolman, and Miller, 1964). Velbel (1985) has asserted on the basis of sediment monitoring in the North Carolina mountains and estimated long-term denudation rates of the Appalachians that 50% to 95% of denudation is produced by high-magnitude, low-frequency events. Jacobson et al. (1989) note that data describing infrequent high magnitude geomorphic events having frequencies of one in 100 years are rare, particularly in steep mountainous

terrain. This paucity of data makes conclusions regarding the frequency-magnitude regime of small catchments tenuous.

In contrast to the idea of geomorphic work dominated by moderate events, Hack and Goodlett (1960) emphasized the importance of low-frequency, high-magnitude events in shaping the Central Appalachian landscape. Wolman and Miller (1960) drew a distinction between the work accomplished by different flow events and the visible change brought about by this work. Wolman and Gerson (1978) elaborated on this idea by defining the term "geomorphic effectiveness" as "...the ability of an event or combination of events to affect the shape or form of the landscape" (p. 190). In this way they related the effectiveness of an event to the rate of restorative processes active on the landscape. In addition, form changes and sediment transport are influenced by the degree of integration of flow processes during an event. For small watersheds, high-magnitude events are more likely to produce greater integration due to expansion of contributing areas throughout the catchment (Wolman and Gerson, 1978). Presumably, integration of process leads to maintenance of form (assuming the process-response system remains fundamentally unchanged), although Wolman and Gerson do not explicitly address this issue. The season of an event also may influence the amount of sediment transported. Wolman (1959) described the seasonal changes in streambank (floodplain) cohesiveness associated with seasonal variations in freeze-thaw activity and soil moisture along Watts Branch in eastern Maryland. Wolman found that bank sediment was most erodible from November through March when freeze-thaw processes were most active.

At a United States Department of Agriculture (USDA) workshop on sediment budgets and routing in forested drainage basins convened in Portland Oregon in 1982 (Swanson et al., 1982), participants addressed the state of frequency-magnitude studies (Kelsey, 1982). Several key issues as well as suggestions for future research were

discussed. Topics included: site- and basin-segment-specific studies, assumption of equilibrium state, climate-specific studies, ergodic reasoning (space-for-time substitution), channel-slope coupling with respect to sediment storage, sediment availability, characteristic frequencies for different processes, event independence, the separation of climatic from geomorphic events, and human influence on frequency-magnitude regimes. Interdependence of successive events in particular was suggested to offer challenges to the application of frequency/magnitude analysis in "steep, geomorphically active, forested terrain" (Kelsey, 1982). Participants concluded that frequency-magnitude analysis had received insufficient emphasis in geomorphology since the pioneering work of Wolman and Miller (1960).

Since 1982, the topic has attracted more attention (e.g., Webb and Walling, 1982; Kochel, 1988; Jacobson et al., 1989; Miller, 1990) and key issues, particularly with regard to catastrophic events and geomorphic effectiveness, are beginning to be explored. In keeping with the issue of site-specific, and basin-segment-specific studies highlighted by Kelsey (1982), Jacobson et al. (1989) note that the Central Appalachian region is one of great physiographic diversity and that the relative importance of different types of events is likely to change across space. Kelsey (1982) concluded that short-term monitoring programs were inadequate for addressing this and other frequency-magnitude issues. To solve this problem, Kelsey (1982) and Jacobson et al. (1989) advocate the maintenance of existing long-term meteorological, hydrological, and sediment transport data collection programs, and the increased use of stratigraphic and dendrochronological methods to extend the analytical timeframe beyond that of direct monitoring.

Anthropogenic Watershed Disturbance and Sediment Yield

Use of land for agriculture, construction, recreation and water supply management affects the operation of processes in the watershed ecosystem, and the resulting system outputs. Much attention has been paid to the role of agriculture in increasing upland erosion and sediment yield, leading to the production of empirical equations such as the Universal Soil Loss Equation (USLE; Wischmeier and Smith, 1965) to predict soil erosion in agricultural areas. Sediment yields from cropland watersheds may be one or two orders of magnitude greater than those from forested but otherwise similar watersheds (Ursic and Dendy, 1965). For the Maryland Piedmont, Wolman (1967) proposed a schematic sequence of sediment yield changes accompanying changes in land use at a fixed location. According to his scheme, a wholly forested area is expected to have sediment yield substantially less than $40 \text{ tons km}^{-2}\text{yr}^{-1}$. With increasing forest clearance and cropping, sediment yield may be expected to increase to around $300 \text{ tons km}^{-2}\text{yr}^{-1}$, an order of magnitude increase over sediment yields under forested conditions. Subsequent reductions in acreage cropped and increase in woodlots and grazing land cause reductions in yields to $80 \text{ to } 150 \text{ tons km}^{-2}\text{yr}^{-1}$. Punctuating this trend may be peaks in sediment yield up to $38,600 \text{ tons km}^{-2}\text{yr}^{-1}$ in zones of construction. After urban development is complete, sediment yields drop to levels similar to or lower than those under forested conditions. Wolman's example reflects very generalized patterns of sediment yield, highlighting the magnitudes of relative differences expected between major land use types. Variation in values of sediment yield across space also occur as a result of the specific nature of land use. For example, Trimble and Lund (1982) documented a tenfold reduction in sediment influx into small agricultural ponds in Wisconsin after the initiation of soil conservation practices (strip cropping and contour plowing) in the 1930s.

In areas having large tracts of forest managed for timber, forestry treatments may be the principal disturbances. For the H. J. Andrews Experimental Forest in Oregon, Frederikson (1970) found that sediment yield from a clearcut basin was 3.3 times that of an adjacent uncut control basin. In a third adjacent watershed with logging roads and a patchwork of cuts, sediment yield was 109 times that of the control basin. For two watersheds in the Coast Range of Oregon, Beschta (1978) found less marked but still large increases in sediment yield after forest treatments. A 25% patchcut with roads and an 82% clearcut with roads produced average annual sediment yields (calculated only over the six years following treatment) of 1.3 and three times (respectively) the values derived from the six years prior to treatment. For upland basins in Scotland, Johnson (1988) found a five-fold increase in sediment yield during the ensuing five years after clearcutting of coniferous forest. For the Coweeta Hydrological Laboratory watersheds in the southern Blue Ridge Mountains of North Carolina, Webster et al. (1988) found elevated levels of suspended inorganic sediment that persisted 30 to 40 years after a variety of clearcut and patchcut activities. Swift (1988) demonstrated for these same basins that forest access roads and skid trails were major sources of sediment from forestry-related activities. The Coweeta sites are among the few where medium-term monitoring data exist from small catchments.

Although sediment yields do often increase dramatically after deforestation as indicated in the above examples, the effects are not always straightforward. Apparently, response to forestry is strongly dependent on specific site and weather conditions and the details of the treatments applied (Binkley and Brown, 1993). In some cases no increases in yield are observed following treatment. Stott (1987) provides an example of the complications that may be found in the relationship between land use and sediment yield. Stott used reservoir deposits from the Macclesfield Forest (Northwest England) lake catchment to document changes in yield over 52 years associated with afforestation (pine

planting) in the catchment. Sediment yields increased instead of declining with forest growth even after canopy closure 20 years after planting. The proportion of subsoil type sediment also increased, apparently because of continued headward growth of deep gullies. In contrast, two adjacent basins experienced none of these trends apparently due to slight differences in land use, vegetation assemblages, and sensitivity of landform and substrate types to erosion processes. Stott's study highlights the need for medium- and long-term data for the purpose of understanding the relationship between land use and sediment yield.

In addition to the destabilization of upland soils and the disruption of the water balance of a catchment, forestry treatments may alter the dynamics of coarse woody debris (CWD) dams which act as temporary sediment storage sites. For undisturbed maturely-forested stream catchments at Coweeta in the North Carolina mountains, Webster et al. (1988) estimate that there are 20 to 30 CWD dams per 100 meters of stream channel. Similar frequencies were also found for the White Mountains of New Hampshire (Likens and Bilby, 1982). Gregory et al. (1993) found frequencies of woody debris dams in the Lymington Basin (UK) similar to those reported in the above cases, and observed that best development of dams occurred in the oldest and least managed areas of deciduous woodland. In the Western Cascades (USA) Nakamura and Swanson (1993) found that CWD dams were the dominant storage controls, particularly in third- and fourth-order basins, and caused widening and local steepening of stream channels. Differences in debris dam frequency between stream orders were linked to size of CWD with respect to channel width, and abundance of CWD was influenced primarily by history of watershed disturbance. Swanson et al. (1982) predict that during the 100 or more years of forest regrowth following logging, old debris dams decay and disappear without replacement since tree mortality is low. The gradual loss of debris dams would result in gradual increase in sediment availability. It is interesting to consider that the

sediment yield effects of historic forestry treatments may have been approximated in the past by natural disturbances like fire, windthrow, and rapid forms of mass wasting. The paucity of medium- and long-term (e.g., successional timescale) sediment monitoring data from small catchments makes the analysis of CWD-sediment yield relationships and forestry impacts on sediment yield problematic. Lake sediment analyses offer one means of expanding the time period for which sediment yield data are available.

Lake Sediment Studies of Sediment Yield

Short-term suspended sediment sampling programs offer the detailed information necessary for event-based explanations of sediment yield and provide initial estimates of medium-term sediment yield. For many small catchments, annual sediment yields are highly variable even in the absence of disturbance (Van Sickle, 1981), making short-term data inappropriate for defining historically and geologically relevant average sediment yields. Extrapolation techniques (e.g., sediment rating curves) involve too much error to be of real use in this regard (Walling, 1988).

One means of obtaining longer records of sediment yield advocated by several authors (Walling and Webb, 1983; Foster et al., 1990; Walling, 1991; Richards, 1993; Walling, 1988) is the analysis of lake sediment. Sediment accumulations in natural lakes often span thousands of years and can be placed in time using isotope dating, biostratigraphy, or less frequently by counting annual sediment laminations found in some lakes (O'Sullivan et al., 1982; Foster et al., 1990). Pollen and other fossil material can be extracted from these sediments to provide synchronous information on vegetation, climate, and disturbance regimes to be matched with estimates of sediment deposition rates (Davis, 1976). Except for lakes having annually laminated sediment, specific years

cannot be firmly placed in the sediment sequence due to the size of uncertainty terms in isotopic dates (usually radiocarbon-based). Even for cases in which individual years can be firmly placed, it remains difficult to single out a cause for a given year's rate of sediment accumulation.

The uppermost sediment in natural lakes, and all sediment in artificial lakes (reservoirs) record trends in both suspended load and bed load components of sediment yield over the medium-term timescale (10^1 - 10^2 years; Dearing et al., 1982; Foster et al., 1988). Younger sediments can often be more precisely dated than older sediments using natural or industrially derived isotopes such as Pb-210 and Cs-137, and occasionally using stratigraphic markers produced by known historical events (Dearing, 1983; Foster et al., 1990). More precise dating of younger sediment allows easier comparison of sedimentological data to historical information for the surrounding environment. For purposes of relating sediment yield to land use, reservoirs have some advantages over natural lakes. First, the beginning of the sediment sequence is precisely dated by the year of reservoir construction. Second, reservoirs may be abundant in areas where natural lakes have not formed. Finally, the watersheds of many natural lakes (e.g., kettle ponds, doline ponds, deflation ponds) often have small surface catchments not formed by fluvial dissection, and are therefore not comparable to the catchments from which monitoring data have been derived.

The USDA has occasionally published the results of reservoir sediment deposition surveys for 1212 mostly large (10s to 1000s of km^2) or fewer small (less than 20 km^2) catchments throughout the US (e.g., Dendy and Champion, 1969). These surveys conducted by various agencies (Bureau of Reclamation, Soil Conservation Service, Army Corp of Engineers, US Geological Survey, Tennessee Valley Authority and others) are most often accomplished by repeated depth sounding (Dendy and Champion, 1969; Dendy et al., 1973), and vary in their accuracy and spatial resolution. Although reservoir

resurvey methods are easy and offer useful results from the perspective of reservoir capacity management, they are limited by the number of years over which repeat surveys are carried out, and are inadequate for documenting actual mineral particulate sediment yields. Documenting only the increasing sediment volume, overlooks the processes of suspended sediment outflow, sediment compaction, organic inputs into sediment, reservoir bank erosion, airborne particulate influx, and autogenic inputs such as the settling of siliceous diatom frustules from the lake water column (Foster et al., 1990). Each of these processes must be accounted for before accurate estimates of mineral particulate sediment yield based on lake sediment can be made.

Techniques of lake sediment analysis have been developed that allow the effects of the above processes to be quantified. For example, McManus and Duck (1985) accounted for compaction, organic inputs, sediment outflow and autogenic inputs in their analyses of reservoir sediment cores from Scotland. The compaction and organic input factors were measured using the simple loss-on-ignition (LOI) procedure of Dean (1974). Autogenic inputs of inorganic material (primarily siliceous diatoms) were estimated by visual inspection with a scanning electron microscope. For measurements of suspended sediment outflow, an estimate of trap efficiency (the proportion of incoming sediment retained behind the reservoir wall) was determined based on the empirical curve of Brune (1953). Brune's curve estimates trap efficiency from the ratio of reservoir capacity to the mean annual water inflow. Other trap efficiency curves have relied on capacity per unit watershed area (Brune and Allen, 1941), and detention time divided by mean velocity of reservoir flow (Churchill, 1948). Based on an enlarged data pool, Heinemann (1981) refined Brune's (1953) curve to require slightly larger reservoir capacity for the same trapping ability. Recognizing that trapping was particle-size dependent, Chen (1975) developed a series of trap efficiency curves corresponding to particle sizes, which were based on the ratio of catchment area to outflow rate. Reservoir bank erosion can be

estimated using erosion pins, although this has seldom been done (Foster et al., 1990). The relative importance of the sediment contribution from bank erosion can be evaluated using the nomogram of Dearing and Foster (1986), which relies on scale and geometric properties of the lake and catchment and estimated regional sediment yield. Airborne particulate influx has received little attention although its contribution to gross sediment accumulation has been estimated to be as much as 9% for a site in England (Foster et al., 1985). This estimate was derived from an airborne-particulate trap located near the lake.

Average mineral particulate sediment yields for the time period following a stream's impoundment can sometimes be further resolved into sediment yields over shorter time intervals. Recent innovation by workers primarily in the United Kingdom has produced new techniques with which to decipher the chronostratigraphy of apparently featureless (massive) lake sediment cores (Foster et al., 1985; Thompson and Oldfield, 1986). These innovations involve the use of bulk sediment magnetic properties to correlate between a network of sediment cores representative of sedimentation across a lake. Once synchronous levels are determined, isotopes can be used to date one or two "master cores" to provide the chronology for deposition across the entire lake. The temporal resolution possible is a function of the accuracy of dating techniques, the temporal distribution of stratigraphic markers, and the depth and degree of bioturbation (mixing by benthic organisms) in the lake sediment with respect to sedimentation rate. Less detailed but longer records of sediment yield may be preferable to those provided by short-term monitoring programs, for questions requiring only average values of sediment yield (Richards, 1993). In addition to the use of magnetic susceptibility as a correlation tool, it is often possible to differentiate sediment sources based on their magnetic signatures (Peart and Walling, 1986; Peart and Walling, 1988). Using sediment mixing models and magnetic characteristics of sediment, a determination of general trends in the relative contributions of different sources may sometimes be possible (Yu and Oldfield,

1993). Although in need of further development, magnetic susceptibility appears to be a useful analytical tool with which to address problems in sediment dynamics over a variety of temporal scales.

Although the potential usefulness of detailed lake sediment analyses for extending sediment yield records has been recognized (Walling, 1988), this approach has seldom been used in the humid eastern US. Two exceptions are notable. Davis (1976) used sedimentation in two Wisconsin kettles to document soil erosion from the surrounding watersheds, giving average rates before and after settlement. Post-settlement change in sediment yield was obtained by assuming constant influx of pollen grains and equating low (diluted) pollen concentrations in cores with periods of increased sediment input from the surrounding upland. The lakes studied are essentially closed basins having no direct stream inflows or outflows so that erosion of slopes bordering the lakes was the only terrestrial source of mineral sediment. In this humid region, high rates of hillslope erosion are primarily related to extent of land disturbance from logging and cropping. The limited size of the surrounding catchment and the lack of fluvial dissection make comparisons of processes and rates of sediment yield with the majority of monitoring studies difficult. The second example is that of Mirror Lake, New Hampshire, near the Hubbard Brook Experimental Forest operated by the United States Forest Service (USFS) (Davis et al., 1985). This study combined ten years of detailed watershed monitoring data and lake sediment analyses at one exhaustively studied site. In contrast to the Wisconsin kettles studied by Davis (1976), the Mirror Lake catchment is drained by three small surface streams which transport eroded soil to the lake. The dominance of Mirror Lake sediment by biogenic material rich in organic matter (gyttja) indicates, however, that lake sediment accumulations are not primarily controlled by fluvial erosion, but instead by autochthonous productivity. Despite the difficulties of comparing sediment yields derived from such a lake to those from most

catchment monitoring studies, research at Mirror Lake stands out as an important contribution to lake-sediment-based sediment yield studies.

Sediment yield studies based on lake sediment stratigraphy are also uncommon from the western United States. A noteworthy exception is provided by Hereford (1987) who studied lake sediments derived from a 2.8 km² semi-arid watershed in southern Utah. These sediments contained a well differentiated visible stratigraphy due to high coarse sediment concentrations typical of flow events in this climate and lack of continual bioturbation in the ephemeral pool. These conditions make small-reservoir-based sediment yield studies in semi-arid landscapes particularly easy.

Issues and Current Problems

The foregoing discussion of sediment yield issues illustrates three important points. First, short-term monitoring alone provides insufficient data for purposes of understanding many key problems in sediment yield studies. These include understanding both temporal and spatial dimensions of sediment yield frequency and magnitude, and clarifying the medium-term responses of sediment yield to natural and anthropogenic environmental changes. Second, lake sediments are a potentially important source of sediment yield information at medium-term timescales. Despite their potential usefulness, detailed lake sediment stratigraphy has rarely been studied in the eastern US with sediment yield problems in mind. Third, it has been asserted by some geographers studying lake sediment-based sediment yield in the United Kingdom that this approach offers a means of obtaining information on the frequency/magnitude characteristics of sediment yield. This assertion has not been substantiated by subsequent research efforts.

The geomorphic work accomplished by high-magnitude discharge events along large alluvial trunk streams is generally understood (Leopold, Wolman and Miller, 1964). However, there are still few data with which to quantify spatial variation in geomorphic work (Jacobson et al., 1989), particularly for small headwater catchments. The ability to study high-magnitude events after their occurrence is increased by the amount of visible form change they often produce. However, large events do not always produce visible form change. Hurricane Agnes in 1972 produced widespread high-magnitude flooding in northern Virginia, Maryland and Pennsylvania (Bailey et al., 1975). Despite reports of unusually high sediment loads, landscape sculpture by this event was virtually unnoticeable in many areas (Moss and Kochel, 1978). Where, within the affected stream basins, were the source areas contributing to observed high sediment loads? How much sediment was contributed by small headwater streams? How do the headwater sediment yields resulting from Hurricane Agnes flooding compare to sediment yields from events of moderate frequency and magnitude in the same basin? Which type of event accounts for more sediment yield over the medium and long term? Is storm rainfall distribution the primary control of event sediment yield, or are other factors like bedrock lithology or drainage basin morphology also important determinants?

The Thompson Lake / Crooked Run catchment in the Blue Ridge Mountains of northern Virginia provides an opportunity to begin answering these questions. This small headwater catchment is located within the zone affected by Hurricane Agnes and possesses the physical and biological characteristics suitable for lake sediment-based sediment yield reconstruction. Analyses of Thompson Lake sediment enable calculation of average medium-term sediment yield over the period since reservoir construction, as well as for decade-scale variations in that average, and for at least one individual high-magnitude event. The geological distinctiveness of the Blue Ridge, the long history of

geomorphic research in the Central Appalachians, the similarity of the site to others (for example, the mountainside opposite the site) experiencing disturbances associated with development, and the proximity of the site to populated areas affected by local water supply issues add to its usefulness.

An analysis of the sediment accumulations in Thompson Lake, and what they reveal about catchment dynamics in this portion of the Blue Ridge Mountains, is presented in succeeding chapters. I begin this analysis by describing in more detail the physical characteristics of the study area and the methods used to derive sediment yield information at the site; these are the topics of the next two chapters.

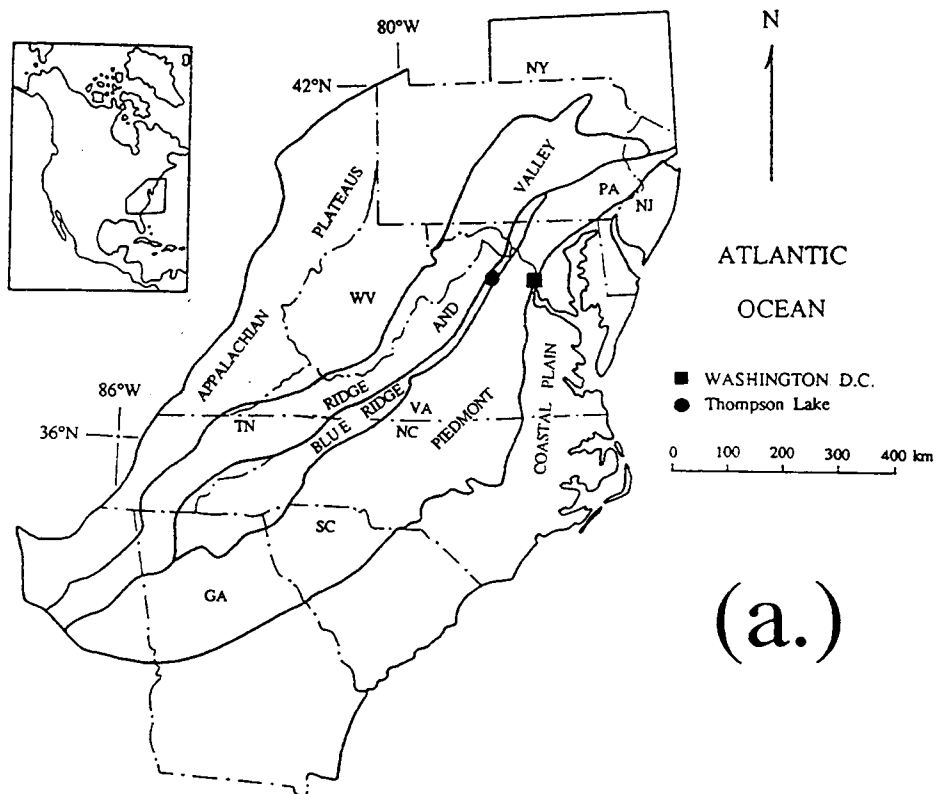
III. ENVIRONMENTAL SETTING

Regional Environment

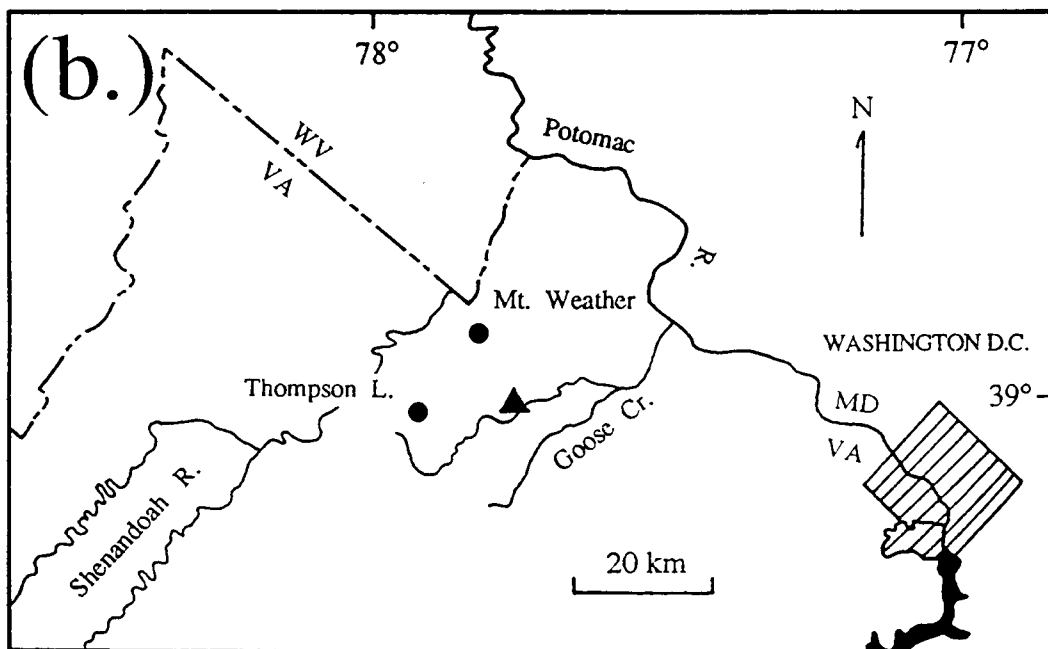
Thompson Lake is located in the Blue Ridge physiographic province of the Appalachian Mountains, a series of ridge-like mountains generally oriented parallel to regional structure (northeast-southwest) and extending from northern Georgia to southeastern Pennsylvania (Fig. 1a). The Blue Ridge is comprised of faulted and folded metamorphic rocks of Precambrian and Paleozoic age (Gathright, 1976). Elevation and width of this province generally decline from southwest to northeast so that at the Virginia-Maryland border the range consists of a single ridge less than 15 km wide and about 600 m high. The western border of the Blue Ridge marks the transition to the Great Valley and the Ridge and Valley physiographic provinces underlain by folded and faulted, but unmetamorphosed sedimentary rocks of Paleozoic age. East of the Blue Ridge is the Piedmont physiographic province which is distinguished from the adjacent mountains primarily by its low elevation and relief rather than its lithology, which is similar to that of the Blue Ridge. The boundary between the Blue Ridge and Piedmont provinces is marked by the Blue Ridge Escarpment, a generally linear and well-defined topographic feature.

Climate of the Blue Ridge is characterized by abundant precipitation (over 100 cm per year in northern Virginia) with the generally low proportion as winter snowfall increasing locally with elevation (Crockett, 1971). Vegetation is composed of a diverse assemblage of broadleaf deciduous trees with local areas of pine (*Pinus* sp.), hemlock (*Tsuga* sp.) and boreal conifers (*Picea rubens*, *Abies fraseri*) determined by disturbance history and topographic location. Inland-moving hurricanes and tropical storms formed over the equatorial Atlantic occasionally track along the East Coast releasing large amounts of precipitation and causing disturbance in the forms of debris flows and river

FIGURE 1. Regional location maps. (a.) The Southern Appalachian region showing major physiographic provinces (adapted from Fig. 1 in Clark, 1989); (b.) Northern Virginia location map showing site locations and the locations of the precipitation and discharge monitoring stations mentioned in the text (adapted from USGS Water Resources data station map; Prugh et al. (1991)). Triangle symbol on Goose Creek denotes the location of the Goose Creek at Middlebury discharge gauging station.



(a.)



flooding (Williams and Guy, 1973; Neary and Swift, 1987). The well-documented effects of Hurricane Camille in 1969 in Nelson County, Virginia include numerous large debris slides primarily derived from topographic depressions (hollows) on hillslopes (Williams and Guy, 1973; Mills, 1987). Mean denudation associated with the combined effects of these slides was from 35.6 to 50.8 mm (Williams and Guy, 1973), a value similar to the long-term Appalachian denudation rate calculated by Hack (1980) of 40 mm per thousand years. Similarities between these values emphasize the potential importance of such events as processes of upland erosion. From sediment monitoring studies at Coweeta Hydrologic Laboratory in the North Carolina Blue Ridge, Velbel (1985) estimates that 50% to 95% of physical erosion occurs via low-frequency, high-magnitude events. Such infrequent events have been considered dominant in long-term denudation (Hack and Goodlett, 1960; Velbel, 1985). Lack of information regarding the long-term importance of events of moderate frequency and magnitude in the Blue Ridge (Mills, 1987) makes acceptance of this conclusion tentative. Clark (1987) has pointed out the difficulty of obtaining information on these infrequently reported smaller events that may be geomorphically effective over the long-term.

Another difficulty in interpreting the relative importance of geomorphic event types in the Blue Ridge is the lack of information on the recurrence intervals of high-magnitude events. Estimates of debris slide recurrence intervals in the Blue Ridge based on those of associated rainfall events range from 50 to greater than 200 years for a major slide event in Pisgah National Forest in North Carolina to greater than 1000 years for the Hurricane Camille event (1969) in the Nelson County, Virginia area (Neary and Swift, 1987; Thompson, 1969). An alternative approach to solving this problem is to date the emplacement of debris flow deposits that form fans along footslope areas (Kochel, 1992). These features are widespread along the Blue Ridge and are variable in their degree of topographic constraint, morphology, composition, soils, and degree of

dissection (Mills, 1987). The preliminary chronology for such deposits in the area affected by Hurricane Camille in Virginia indicates a recurrence interval of approximately 3000 years for such an event (Kochel, 1992). These techniques are of less use for more topographically constrained fans, commonly flooring basins in steep terrain, which presumably are destroyed by erosion between events (Mills, 1987). Data on processes and rates of fan removal in constricted fans, factoring in the effects of historic land use, are not available to test this presumption.

The degree of adjustment between Blue Ridge landscapes and the current climate has been a topic of debate. Hack (1965) first proposed that Central Appalachian landscapes were in a state of dynamic equilibrium in which landforms were primarily adjusted to structure, and did not undergo appreciable form change in the absence of structural change (e.g., exposure of new structure by downwasting). That the degree of adjustment is not full is evidenced by the occurrence of many relict periglacial landforms on and marginal to Blue Ridge summits as far south as North Carolina (Clark and Ciolkosz, 1988). Considering that warm interglacial conditions have probably been present a small proportion of the Pleistocene relative to glacial conditions, Braun (1989) has proposed that over geological time, higher Appalachian areas may be more closely adjusted to glacial period climate than to the current climate. From the standpoint of current landscape stability, the interaction between these relict features and current climatic conditions is an important research topic. For example, Whittecar and Ryter (1992) have suggested that the high permeability of relict periglacial boulder accumulations in first-order drainages reduces the likelihood of debris avalanches.

Site Description

Thompson Lake (38°57'25"N, 77°59'40"W; Fig. 2a and 2b) is a small (3.22 ha) reservoir located on the eastern flank of the Blue Ridge Mountains in northern Virginia at an elevation of 293 m. The lake receives the drainage of a 3.83 km² second-order catchment (Crooked Run) heading near the crest of the Blue Ridge at 640 m elevation. The impounding structure was built after April, 1965 and was completely filled by October, 1966 according to aerial photography, giving a reservoir age of 30 to 31 years. The structure consists of an earthen dam with grass-floored emergency spillway and a principal spillway consisting of a vertical riser. It is not known if the emergency spillway has ever been used; fluctuations in water level greater than 30 cm are apparently rare (Irving Kenyon, former state land manager for the region, personal communication, 1994). Normal capacity for the reservoir is estimated from bathymetry to be 116,340 m³ (top of principal spillway) and maximum capacity according to a 1974 engineering inspection report is 370,302 m³ (emergency spillway level); maximum depth is around 9 m. Small-scale mapping of runoff regimes (Riggs and Harvey, 1990) suggest an estimated annual inflow from the upstream watershed of 1,900,000 m³. Occasional direct measurements of surface inflow rates I have obtained during each season are lower, but generally consistent with this estimate. Comparisons with data from Brune (1953) and Heineman (1981) suggest that these volumes should allow for trap efficiency of 81% for the normal pool volume. This evaluation assumes that inflow of groundwater into the reservoir is balanced by groundwater outflow through the earthen impounding structure. Unlike many other reservoirs of similar age in the region, Thompson Lake has neither been drained nor dredged so that the record of sedimentation remains intact.

The catchment bedrock consists primarily of highly resistant Catocin Formation greenstone (Gathright, 1976). The second-order stream channel erodes a topographically-

FIGURE 2. The Thompson Lake/Crooked Run catchment. (a.) Topographic map of Crooked Run above Thompson Lake (outlined perimeter) and Wildcat Hollow basin (see Methods, chapter 4) to the south. (from: USGS 7.5' topographic quadrangles; Linden and Upperville, Virginia, 1990); (b.) Thompson Lake and immediate vicinity. Small dark bars with "L#" designation mark locations of erosion pins in lake banks (see Methods, chapter 4).

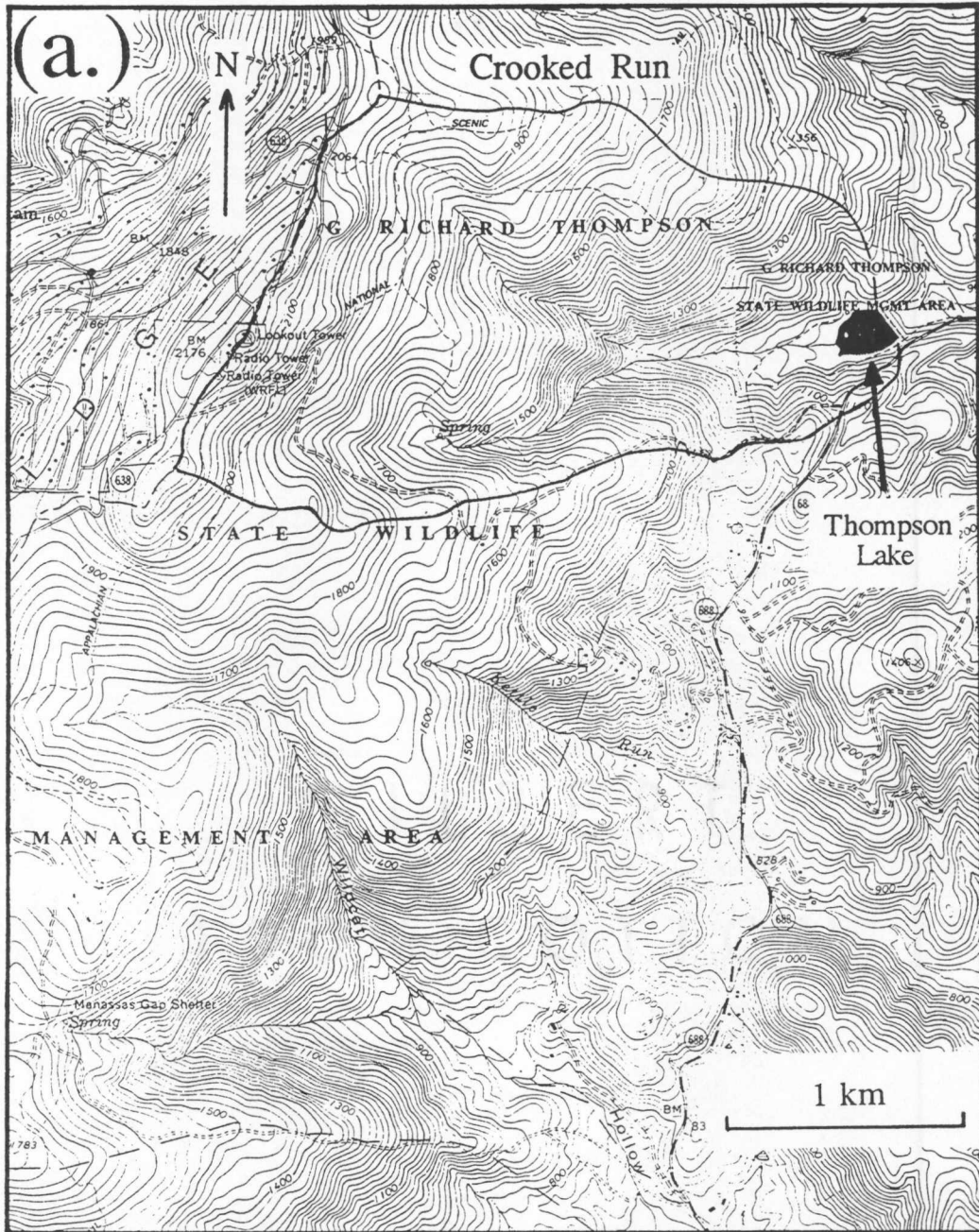
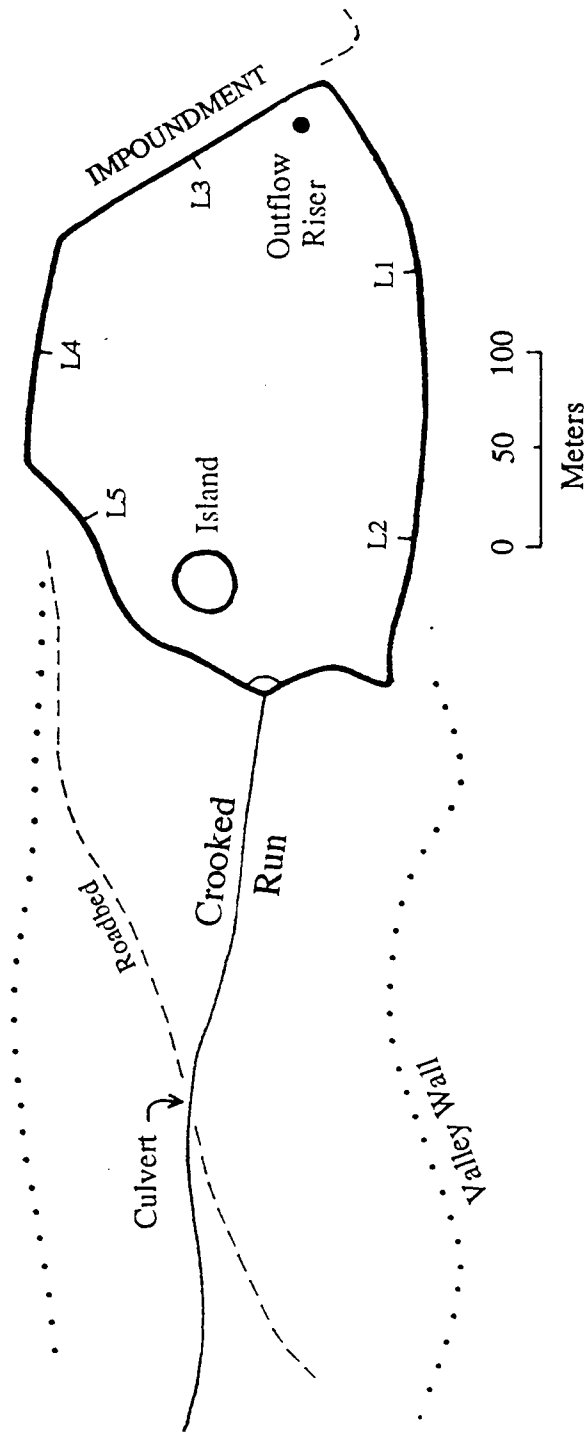
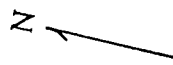


FIGURE 2 (continued)

(b.)

Thompson Lake

Impoundment : EARTHEN
 Area : 3.22 ha
 Normal Pool : 116,340 m³
 Perimeter : 767 m
 Age : 30 y
 Trap Eff. : 81%



constrained debris fan or "train" surface which has been fluvially incised as deeply as 3 m; average depth of incision decreases downstream. Bare channel walls expose highly weathered subsoil in the fan to active erosion evidenced by bank collapse, tree root exposure and widespread tree topple, and micro-pedestal formation associated with raindrop impact. The geological characteristics of the Thompson Lake watershed are replicated in nearby basins extending along the Blue Ridge Escarpment north of Manassas Gap (4.8 km south).

Judging from nearby forests of known age, the basin appears to have been widely logged around 1940. Forests of similar stature in adjacent catchments to the north are judged to be 30 to 80 years old (personal communication, Sky Meadows State Park Staff, Virginia). Air photos show selective cutting in portions of the higher elevations in 1965, and one small patch of forest cleared in the north headwater during or prior to 1990. At least three homes existed in the basin during the 19th century, one on the fan surface and two in the northern headwaters (Virginia Outdoors Foundation, 1990). No information is available on the extent of farming and grazing associated with this settlement. The boulder mantle over much of the second-order valley may have confined prior agricultural activities to the higher uplands which are less steep than sideslopes of the valley walls.

Today the basin is occupied by a variety of deciduous tree types including sycamore (*Platanus occidentalis*), ash (*Fraxinus*), maple (*Acer*), basswood (*Tilia americana*), and beech (*Fagus grandifolia*), some of which are large, particularly on high stream banks. The large size of trees along some stream reaches may indicate incomplete clearance of these zones during logging. The northern-aspect valley wall bordering the lower portion of the second-order valley supports younger forest in former areas of orchard shown on topographic maps as late as 1970. Aerial photos, however, show these areas to be mostly forested by 1965. The valley wall opposite this location was also in orchard in the 1940s.

The closest source of detailed meteorological data over the past 50 years is the Mount Weather meteorological station located 14.5 km northeast of Thompson Lake along the Blue Ridge crest (Fig. 1b). Although not directly recording weather events at Thompson Lake, this station likely experiences annual distributions and amounts of precipitation similar to those at the study site because of its proximity and similar topographic location. Major frontal rainfalls and inland-moving tropical cyclones affect broad areas and can be identified for Thompson Lake by cross-referencing daily precipitation recorded at Mount Weather with surrounding stations. In the absence of direct monitoring, extrapolation of general conditions at Mount Weather gives the best available weather record for Thompson Lake.

Precipitation regimes over the last 30 years at Mt. Weather range from near average annual totals (~100 cm) during the 1960s, to above average totals with several large storms during the 1970s (NOAA, Climatological Data, Virginia). Annual precipitation was often lower than the average value of 100 cm, during the early and mid-1980s, and average or slightly above in the early 1990s. Rainfall events associated with well-documented storms are recorded for this and surrounding stations; Hurricane Agnes (June 22, 1972; 19.4 cm), October, 1976 storm (water year 1977; 10.5 cm), August 24, 1979 (14.1 cm), and November 4th, 1985 storm (part of Tropical Storm Juan; 24-hour rainfall of 5.2 cm). The Hurricane Agnes rainfall event which produced the highest of these totals represents an unusually high-magnitude event of low frequency (recurrence interval greater than 100 years; Bailey et al., 1975) and large spatial extent (pre-landfall storm diameter of 1600 km).

No stream discharge data exist for the Thompson Lake catchment. The nearest source of discharge information for watersheds draining the Thompson Lake area is located at the Goose Creek near Middlebury station measuring discharge from a 300 km² catchment, including the study site.

IV. METHODS

Orientation

In this methods chapter I describe the acquisition of information needed to reconstruct and interpret sediment yield from the Thompson Lake catchment. This information falls into the categories of watershed morphology and processes, environmental history, and reconstructing sediment yield from lake sediment deposits. Information on the first two categories is required for interpreting the sediment yield reconstruction. The sediment yield reconstruction is based on lake sediment stratigraphy defined by mineralogical and isotopic changes through a series of sediment cores. The methods described in the following two sections are oriented accordingly.

Field Survey and Basin Monitoring

Basic field mapping and process monitoring within the watershed allow the qualitative assessment of sediment sources and provide information useful for interpreting lake sediment accumulations. I mapped the channel morphology of the second-order portion of the stream network, and noted cultural features such as old roadbeds and rock walls indicative of former land use or habitation. The first-order streams of the catchment do not flow in discrete channels, but instead through wide boulder-mantled debris accumulations. Mapping was accomplished through meter tape and compass survey along the channel, with notations regarding approximate width and morphology of channel cross sections. Accumulations of coarse woody debris within the second-order channel were mapped and characterized with regard to the size (diameter) of fragments and their

state of decay. The determination of decay state was based on the classification scheme of Nakamura and Swanson (1993). Debris dams are potentially important for medium-term sediment storage dynamics, and provide a link between sediment yield and forest stature.

Initial reconnaissance indicated the potential importance of streambank erosion as a means of mineral sediment input into the fluvial system. I placed 15 cm-long x 7 mm-diameter steel erosion pins (painted fluorescent orange) in streambanks at ten locations along the channel. Channel banks greater than 0.5 m in height received two or more pins distributed vertically, but slightly offset (horizontally) up the bank. Pins were spaced a minimum of 50 cm apart and were left sticking out 3 cm. All pins were driven into the bank with a hammer; the generally cohesive bank material remained intact as pins were driven in. Changes in exposed pin length were measured at succeeding visits to the field area. No evidence of enhanced erosion or deposition in the immediate vicinity of the pins was observed, and bank recession appeared to be uniform where it occurred. These measurements are intended to reveal the general level of streambank erosion at the site. More important for interpreting lake sediment accumulation is the erosion rate of the lake shoreline, which introduces non-catchment-derived mineral sediment into the lake. Few data are available on this form of sediment input at this or any other site (Foster et al., 1990). I placed erosion pins in lake banks at four locations along the lake shore (Fig. 2b), and measured their exposure with time to estimate the rate of bank recession for the ten sites.

To gain an informed perspective on annual flow volume, current sediment yields, and the hydrologic response of Crooked Run, I measured discharge and sampled suspended sediment at each visit to the site. The frequency of discharge measurements varied from one measurement every two days during dry periods, to several measurements during a two-day precipitation event. Suspended sediment was not sampled at every discharge measurement; more samples were taken during precipitation

events than during baseflow conditions. Ordinarily, I measured discharge by averaging the result of three direct flow collections at a culvert just upstream of the lake. For some events, I measured discharge by estimating average flow velocity from floating object velocity and multiplying times the cross-sectional area of flow (i.e., the velocity-area technique). I also occasionally used the geometry of the culvert, along with depth measurements of culvert flow to estimate velocity and discharge using the Manning Equation. During floods, direct flow collection was not possible, making these indirect techniques the only ones feasible. In association with discharge monitoring, I used rain gauges to estimate concurrent precipitation. Gauges were placed by the lake, and when possible, in the headwaters within a cleared zone bordering the ridge crest road (Fig 2). These monitoring activities led to the collection of ten suspended-sediment samples, 16 discharge measurements, and measurements of associated precipitation, over the course of two years. These measurements include data for one moderate flood event in the summer of 1995, and are representative of seasonal contrasts in antecedent moisture. Suspended-sediment samples were collected by hand in one-liter bottles, centrifuged at 2000 RPM for 14 minutes in a large centrifuge, and transferred to crucibles for loss-on-ignition analysis.

Although suspended sediment and bed load from the stream channel probably represent the most important sediment sources for Thompson Lake, input of airborne mineral particulates has been found to be important at some other sites (Foster et al., 1990). Because of the fully-vegetated landscape surrounding Thompson Lake, I expect airborne particulate input at this site to be minimal. As a general test of this assumption, I collected fallout particles by the lakeside in a 28-cm-diameter plastic bucket partially filled with glass beads to trap particles (Goudie, 1981). Three small drainage holes near the top of the bucket allow excess rainfall to flow out with minimal loss of trapped particles and the top was covered with 0.6 cm (diameter opening) steel mesh to keep out large organic

debris. I collected the washings from this bucket after six months and 1.5 years, centrifuged them for 14 minutes at 2000 RPM, and transferred the pellet to crucibles for loss-on-ignition ("LOI"; Dean, 1974) to isolate the mineral fraction. The total mass of mineral particles was divided by the collecting area of the bucket and the time of collection to estimate fallout contributions to lake sediment.

Environmental History

Interpretation of lake sediment sequences requires information on the state of the watershed environment both before and after the lake's impoundment. Because no monitoring data are available from within the study catchment, information about past environmental state must be derived from other sources. These include anecdotal information, aerial photography, early map editions, discharge and precipitation data from stations nearest to the site, and dendrohydrology. I gathered anecdotal information through telephone and personal conversations with residents and government personnel. I acquired aerial photography from the United States Geological Survey (USGS) showing the catchment in 1965, 1969, and 1990. This sequence of photos provides information on major land use and forest stature changes over the 29 years since reservoir construction. History of habitation near the site was derived from Virginia Outdoors Foundation maps depicting 19th century home sites within the catchment. A summary of information from these sources was presented in the preceding chapter.

Discharge and precipitation records are of critical importance primarily for the period following construction of the impoundment. I obtained files of daily (and for some days, 15-minute) precipitation at Mt. Weather, 14.5 km northeast along the Blue Ridge crest (Fig. 1b), for the past 40 years. I obtained similar daily records of discharge from

the Goose Creek at Middlebury station (Fig. 1b), the nearest station incorporating the Crooked Run drainage. These data were used to approximate the chronology of past floods and their relative sizes in Crooked Run. Individual precipitation events suggested by these data were cross-checked with other stations in the region to determine which events may have been regional in spatial extent. Largest discharge events in Goose Creek probably reflect widespread precipitation in the study area and therefore were also of value in identifying events in Crooked Run.

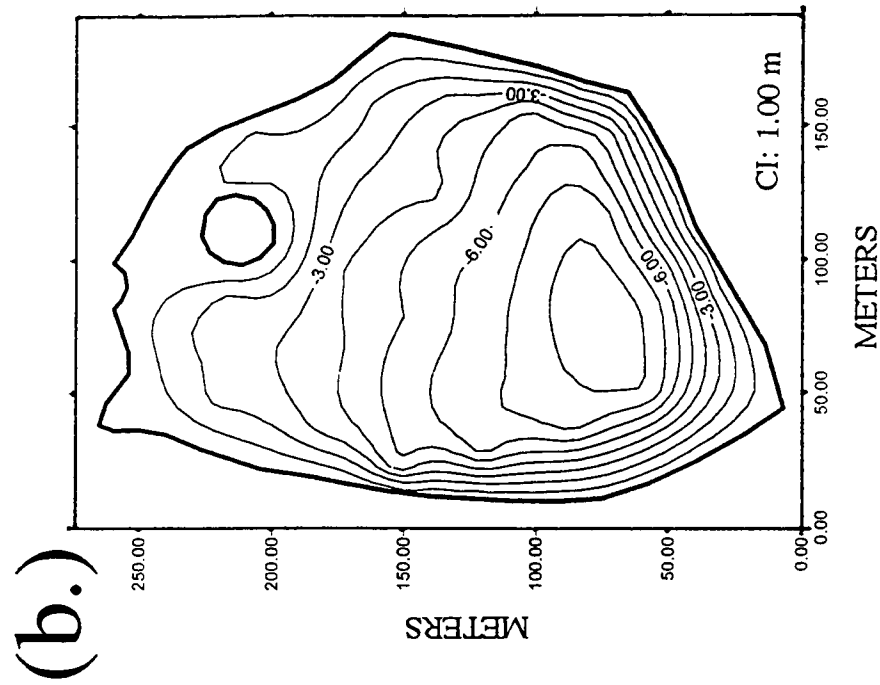
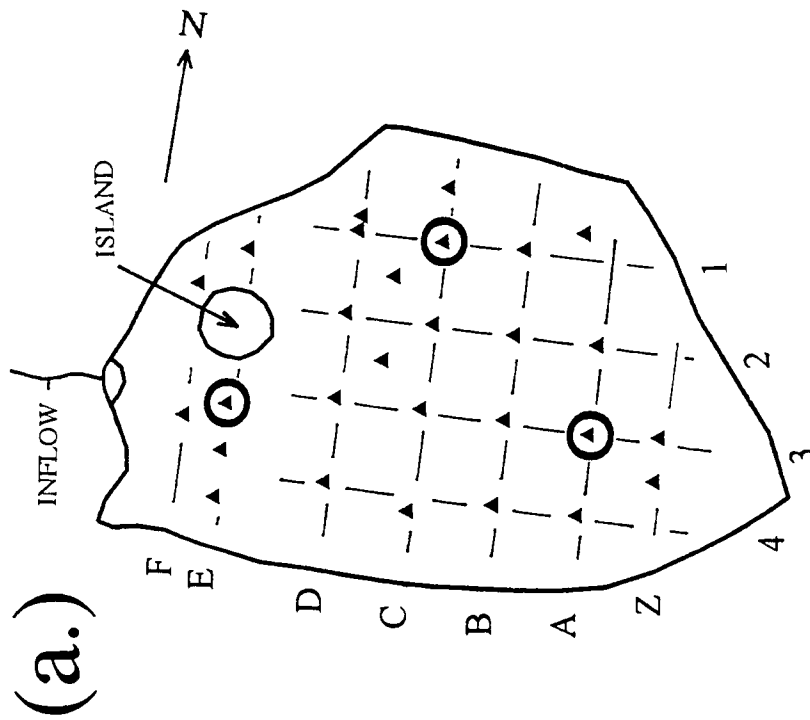
To corroborate inferences of flood chronology from discharge and precipitation data, I dated potential flood scars on trees growing along the second-order channel of Crooked Run above Thompson Lake and along Wildcat Hollow, a similar basin located just southwest of the study site (Fig. 2). Studies of tree-scar ages on the upstream sides of riparian trees have proven useful for documenting flood dates along large channels, but have seldom been used to study flooding along smaller streams. In most cases, wedges were recovered from tree scars, clearly showing the number of annual growth increments since damage occurred. When this was not possible, I extracted increment cores from the scar face and a bole position outside the healed area. All wedges and cores were dried and sanded with 400-grade sandpaper. Skeleton plots of annual rings in increment cores were used for correlating ring sequences from the same tree and to compare ring width variation between trees. Formal cross-dating, in which a sequence of ring widths from a tree is compared to several other independent sequences to provide an absolute date, was not possible. Instead, tree-ring dates in this study were derived from simple ring counting, sometimes used in dating scars (Phipps, 1985). Sprouts clearly associated with tree tilting or toppling along stream banks were also sampled; these were sectioned, sanded, and counted to provide estimates of the age of disturbance. Unlike tree scars, unique ages often cannot be established for sprouts because of variable delay in sprouting following disturbance. Tree species information and precise locations for each tree

sampled were recorded. In addition to scar and sprout ages, I noted the ages of thin-ring sequences which may be produced by lesser damage to tree bark than is responsible for tree scars (Alestalo, 1971). In some instances, I was able to determine tree establishment age which is potentially indicative of vegetation destruction or creation of new alluvial substrate by flooding. I interpreted the coincidence of two or more occurrences of this dendrohydrological evidence to constitute a potential flood event.

Lake Sediment Analyses

The sediment yield reconstruction requires the delineation and correlation of lake sediment strata across the lake, and the placement of absolute ages on these strata. This is accomplished by measuring various physical and chemical attributes of lake sediment cores. I collected sediment cores in 5.8-cm-diameter plexiglass tubes at 28 locations across the lake surface (Fig. 3). Rubber pistons within the plexiglass tubes were used to retain the loose lake sediment within the tubes during retrieval. Coring locations (Fig. 3a) correspond to intersections on a grid with cell dimensions of 35 x 35 meters in the deep central portion of the lake (Fig. 3b), and to locations along radial transects extending from the island to shore in the deltaic region. A few supplemental cores between grid intersections were obtained for better coverage of key areas. This distribution of grid points samples all portions of the lake without requiring that a large number of cores be retrieved. Additional information on the thickness of sediment accumulations in the shallow deltaic region was derived from sediment probing with PVC (poly-vinyl-chloride) tubing. Three of the coring sites (A3 from the deep water region, C1 from moderate depth, and E3.2 from the delta) represent "master core" locations where duplicate cores were retrieved to provide material for multiple analyses (Fig. 3). Coring

FIGURE 3. Thompson Lake coring grid and bathymetry. (a.) Coring grid; triangles represent coring locations and circled triangles show master core locations; (b.) Bathymetric map; isobaths are in meters and contour interval is one meter.



was accomplished during three visits: October, 1994 (5 coring sites), June, 1995 (12 coring sites), and December, 1995 (11 coring sites). Two additional sites were cored in July, 1996 to aid in correlation between the main grid points. The time spread for core collection introduces some error into sediment yield calculations for recent sedimentation. To map the lake's bathymetry, routine water depth measurements were made at each grid intersection and at midpoints between them along north-south transects of the grid. Mapping bathymetry is important for interpretations of uniformity in sediment accumulation. All cores were described and transported upright (unextruded) to the University of Tennessee for subsampling. I extruded all cores at 1-cm increments to obtain a series of 1-cm-thick slices, and stored these slices in plastic "ziploc" baggies in a cold room maintained at a temperature of 8°C.

The properties selected for defining the lake sediment stratigraphy are loss-on-ignition (LOI)(Dean, 1974), magnetic susceptibility, and levels of the radionuclides Cs-137 and Pb-210. Loss-on-ignition analyses give values for organic matter content and bulk density that are used to convert sediment volume calculated for a given stratum into total mass of mineral sediment. Magnetic susceptibility is a simple, rapid and non-destructive measurement of how magnetizable a material is. It reflects the mineralogical and particle-size characteristics of sediment and has been proven useful as a correlation tool (Foster et al., 1990). Measurements of Cs-137 and Pb-210 activity provide a means of constructing an absolute chronology for the deposition of sediment strata defined by the magnetic profiles. Cesium-137, an artificial radionuclide produced by industrial nuclear testing, accumulates in soils and sediments by fallout in proportion to its monitored atmospheric concentration (Ritchie et al., 1973). Lead-210 is a naturally occurring radionuclide having a decay rate that makes it useful for dating recent sediment. The details of its use are more fully laid out in chapter 7 and the Appendix.

I measured loss-on-ignition and magnetic susceptibility at one-centimeter intervals for each of the three master cores. Cesium-137 was measured at one-centimeter intervals in master cores A3 and C1, and Pb-210 was measured at one-centimeter intervals at site A3 only. These measurements were obtained using gamma-ray spectrometry on 15 cm³ of wet sediment packed into sterile plastic petri dishes. Gamma counting was accomplished at the Environmental Sciences Division, Oak Ridge National Laboratory on a Nuclear Data 6000 spectrometer. The determination of Pb-210 content from its low-energy gamma requires specialized techniques to correct for absorption effects of the sediment matrix. Lead-210 counting was done by I. L. Larsen, Environmental Science Division, using the counting technique of Cutshall et al. (1983). After counting, all samples were air-dried and weighed to allow mass-specific calculations of radionuclide content.

Magnetic susceptibility was measured on air-dried one-centimeter-thick core slices for all cores. Samples from radionuclide analyses were used for magnetic susceptibility determinations on master cores. I crushed each dry sample using mortar and pestle and packed the crushed material into a 10 cm³ plastic cylindrical pot. Each pot was weighed before and after packing to enable calculations of dry mass and mass-specific magnetic measurements. I measured volume susceptibility (k) at both low and high frequencies using the Bartington Instruments MS2 meter and MS2B dual-frequency sensor. These measurements allow the calculation of mass-specific low frequency magnetic susceptibility and the frequency dependence of susceptibility. After analysis, dry samples were stored in 30 ml plastic vials for possible future measurements. I used the resulting magnetic profiles from the sediment cores to correlate sediment sequences from different parts of the lake. I correlated the cores by qualitative means through visual comparison of the magnetic profiles. This approach, although subjective, allows consideration of the observer's experience with other properties of the sediment cores. Further justification

based on the actual data is given in chapter 7. I also measured the magnetic susceptibility of nine bulk samples from potential sediment source areas of the channel bank (subsoil and topsoil) and upland hillslope topsoil, as well as from sediment samples collected directly from within the channel.

I measured the particle-size distributions in portions of one master, and two other cores, using the pipette method (Gee and Bauder, 1986) to aid the interpretation of the magnetic data. Meeting sample-size requirements for the pipette procedure sometimes necessitated the combination of three (successive) one-centimeter slices, strategically located according to important variations in magnetic susceptibility. Magnetic susceptibility of samples was measured before particle-size analysis so that chemical dispersants would not affect susceptibility values. Particle-size analyses are important for interpretation of susceptibility values which are dependent on magnetic mineral size.

Adjustments to calculated sediment strata masses must be made for losses through outflow of suspended sediment in the lake water column before deposition (i.e., trap efficiency), and additions of sediment from airborne fallout, sedimentation of biogenic silica from the water column, and lake bank erosion. I estimated trap efficiency using Heineman's (1981) curve requiring estimates of water residence time in the reservoir. I estimated this residence time by comparing mapped values of average annual runoff to the measured reservoir volume. (As stated in the prior chapter, the resulting trap efficiency was 81%.) Airborne fallout was measured using the particle trap described earlier. To determine biogenic silica (mainly diatom) content of lake sediment, I mounted LOI residues (after 550 °C) on glass slides for microscopic examination. Observations of LOI residues from organic-rich sediment retrieved from sites in New York (see Clark et al., 1996) show that diatom frustules survive this procedure and therefore can be used to indicate relative abundance. I only found two diatom frustules even after repeated scans of entire cover slip areas, suggesting that this input is unimportant at Thompson Lake.

The results of applying the methods outlined in this chapter are presented in the following three chapters in order of their treatment here. Chapter titles correspond accordingly to each of the preceding sub-section headings in this chapter.

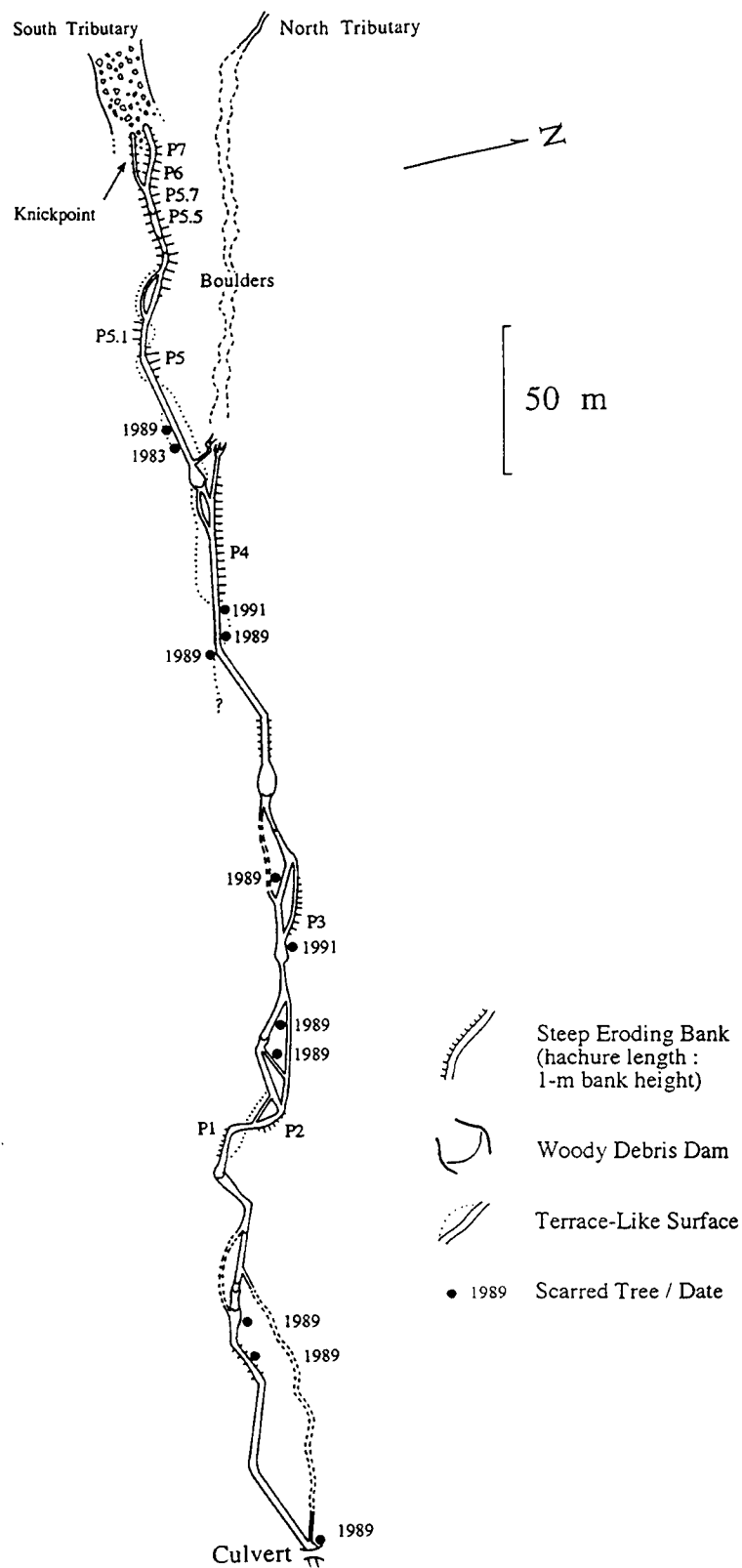
V. FIELD SURVEY AND BASIN MONITORING

Field Survey

Figure 4 shows the channel morphology of the second order portion of Crooked Run, and also serves as a location map for erosion pins and tree-ring samples. I did not map the portion downstream of the culvert shown on Figure 2b. This segment continues the morphology exhibited just upstream of the culvert (flow between low banks) and is surrounded by dense thickets of greenbriar. The total length of mapped channel is 550 m. The additional length downstream of the culvert brings the total length of the second-order portion of the catchment to around 700 m. Along most mapped segments, one main channel is clearly visible. In some stream reaches however, segments bifurcate leading to alternative channels that are inundated with only slight elevation of the stream surface. These secondary channels rejoin the main channel a short distance downstream.

Incision of Crooked Run into the underlying debris train within the second-order valley ranges from 3 m in the upper half of the mapped area to no incision in the lower half. The general decline in incision depth is not constant downstream; incised stream reaches are interrupted by short segments with low or no banks where water and sediment may leave the channel during flooding. This process is evidenced by scoured areas roughly parallel to the main channel that sometimes rejoin it through small feed-in gullies. The upstream end of the mapped second-order channel is marked by a clearly defined knickpoint 0.5 m in height. This feature is located just upstream (along the south tributary) of the stream junction formed by unchannelized flow from the north tributary meeting the main channel and creating two deep gullies (Fig. 4). Subsurface flow derived from the north tributary causes continuous saturation of the channel banks downstream of the knickpoint leading to frequent bank collapse along the main channel.

FIGURE 4. Map of the second-order channel upstream of Thompson Lake. Locations for erosion pins are designated "P#", and locations for tree scar samples are designated by dots with year of scarring attached. The map legend shows selected morphometric properties depicted on the map. (Survey by author).



Dotted lines along the main channel shown on Figure 4 mark the locations of small terrace-like landforms. Two surfaces (not differentiated on the figure) are often discernible; a low (15-30 cm above stream) and usually narrow (0.5-2 m wide) surface that is inset along the main channel into a higher (~1 m) and wider (> 2 m) surface. The zones are rarely continuous on both sides of the channel, but sometimes form a band alternating from side to side downstream. The lower surface showed signs of inundation by a moderate flood in the summer of 1995 (see page 53 for more on this storm). The higher surfaces also show signs of past fluvial influence such as occasional small chute-like landforms along their distal (away from the channel) edges. Evidence for frequent inundation is lacking however, and their origin may also involve mass wasting (bank collapse), or earlier debris damming and aggradation.

Hupp and Osterkamp (1985) have noted similar features along other small streams in Virginia and have drawn inferences regarding their hydrological nature. Three surfaces commonly identified by these authors along first- and second-order streams are depositional bars, channel shelves and floodplains. Depositional bars are deposits along channel margins that are inundated about 40% of the time (Hupp, 1988) and are usually occupied only by herbaceous species. Channel shelves occur in areas between depositional bars and the channel banks (cut into the floodplain) that are inundated 10% to 25% of the time; typical vegetation on these landforms includes alder (*Alnus serrulata*), willow (*Salix nigra*) and arrow wood (*Viburnum dentatum*). Floodplains are inundated with a recurrence interval of one to three years and support a number of trees including (prominently) American elm (*Ulmus americana*), bitternut (*Carya cordiformis*) and black walnut (*Juglans nigra*). Depositional surfaces lower than the lowest surface included on Figure 4 at Crooked Run support abundant alder and are therefore similar in vegetation to the shelf environment. The lowest surface included in the mapping of terrace-like landforms on Figure 4 is usually narrow and inset into the higher surface, and often lacks

woody vegetation, perhaps in part due to this small size. In places where this surface is vegetated, ironwood (*Carpinus caroliniana*), a plant more typical of the floodplain environment, but also found on channel shelves (Hupp, 1988), is common. The higher surface supports a number of woody species including ironwood and elm. Judging from observations of the stream channel in flood, the lowest surface included in Figure 4 is more likely floodplain than shelf; the higher surface, if in fact alluvial in origin, may be a rarely inundated terrace despite the bottomland type species sometimes found growing there. This cool riparian environment is usually moist even during drought.

Erosion Pins

Locations on Figure 4 marked "P#" designate locations of erosion pins; erosion pin data are presented in Table 1. This table shows the amount of new exposure of erosion pins produced by channel bank erosion during time intervals between visits to the site. Pin number designation increases with distance upstream from the culvert. Approximate bank heights for erosion pin sites are indicated by the length of channel bank hachures on the channel map (Fig. 4). Maximum values for erosion pin exposure are in excess of 20 cm from the dates 2-20-95 to 12-3-95. Most of this erosion may have resulted from a wet period during the third week of June 1995 that ended with a large storm on June 21. Five pins were removed by erosion between 2-20-95 and 8-2-95. Pin removal was by bank collapse, and large portions of collapsed bank lay *in situ* on 8-2-95 even after the June 21 storm. Areas of bank collapse all lie within upper reaches of the second-order channel where active incision continues, and where banks remain saturated due to constant input of subsurface waters from the north tributary. Other high values of retreat occur along portions of the channel where streamflow is concentrated, such as in sharp bends (P1,

TABLE 1. Erosion Pin Data.

Pin #	New Exposure 8-6-95 (cm)	New Exposure 12-3-95 (cm)	New Exposure 7-29-96 (cm)	Total (cm)	Notes
P1a	5.0	0.0	9.0	14.0	large storm, 6-21-95
b	5.0	0.0	9.0	14.0	
P2a	2.0	0.0	- 2.0	0.0	
b	6.0	0.0	- 1.0	5.0	lost?
P3a	1.0	0.0	1.0	2.0	
b	6.0	0.0	NL	6.0	
P4a	- 1.0	0.0	0.0	1.0	
b	NL	NL	NL	NL	
P5a	- 1.0	1.0	NL	0.0	
b	- 1.0	1.0	NL	0.0	
c	- 1.0	NL	NL	- 1.0	
P51a	1.5	- 1.5	4.0	4.0	
b	1.5	- 1.5	NL	0.0	
P55	5.5	removed	NL	>19	large storm, 6-21-95
P57a	3.0	- 0.5	1.5	4.0	
b	1.0	- 1.0	6.0	6.0	
P6a	removed	---	---	>15	
b	removed	---	---	>15	
P7a	removed	---	---	>15	
b	removed	---	---	>15	
c	removed	---	---	>15	lost ?
L1	0.0	0.0	- 0.5	- 0.5	
L2	NL	NL	NL	NL	
L3	2.0	NL	3.0	5.0	
L4	1.0	NL	NL	1.0	
L5	NL	NL	NL	NL	lost ?

NOTE: All pins emplaced 2-20-95. Pins with the "P" designations are located in streambanks; those with the "L" designation are located in the reservoir banks. Exact locations are shown in Fig. 2. Notations: "Removed"- pin removed by erosion; "---" - removed pin not replaced; "NL" - pin not located.

P2) and to the sides of mid-stream boulder or stable debris accumulations (P5.5). During fall, 1995, I recorded little change in bank erosion. Values of new pin exposure were usually zero or even negative during this relatively dry period, reflecting net accumulation of sediment from upslope or *in situ* loosening or expansion of soil due to turbation processes (e.g., freeze-thaw) during the late fall. These data emphasize the importance of rainfall events for channel erosion and also point to the potential importance of streambank conditioning in making sediment more erodible during intervening dry periods.

Erosion pins installed in the lake banks (locations on Fig. 2b) indicate far less erosion than do those along the channel. Data from these pins, however, are limited due in part to difficulties relocating them during the winter visit of 12-3-95. For all erosion pins throughout the catchment, exposure values of ± 0.5 cm are difficult to interpret because of the variability associated with changes in measuring conditions between visits. A maximum retreat of five centimeters occurred for pin L3, whereas much less retreat occurred for pins L1 and L4. Pins L2 and L5 have not been found again since emplacement and are judged to have been covered by debris rather than removed by erosion, since 15 cm of retreat would have an observable effect on shoreline vegetation. No evidence of such change was observable at these pin sites.

The average bank retreat value derived by assuming that the hidden pins represent zero retreat over 1.42 years equals 0.84 cm yr^{-1} . The short duration of pin monitoring makes this value tenuous. An additional difficulty in using erosion pin data in this study is potential change in bank retreat rate through time. Highest bank retreat rates may have occurred just after initial reservoir filling. Over what specific time periods current retreat rates apply is unknown.

Precipitation, Discharge, and Suspended Sediment Concentrations

The measurement dates, times and values for precipitation, discharge, and suspended sediment monitoring are presented in Table 2. Lowest values for baseflow discharge ($<10 \text{ l s}^{-1}$) occurred during late summer and early fall, and highest values ($\sim 35 \text{ l s}^{-1}$) during winter and spring. Extrapolated over six months and added, these flows amount to an annual baseflow discharge of $710,000 \text{ m}^3$. Note that this value is far less than the $1,900,000 \text{ m}^3$ predicted from regional runoff mapping (Riggs and Harvey, 1990). The best estimate of the average value may lie somewhere in between these two. The higher value was used in the estimation of trap efficiency; trap efficiency varies inversely with annual discharge for a given lake volume (Heineman, 1981). Discharge increased for all but the smallest precipitation events monitored or noted. The most marked change occurred in association with heavy rainfall beginning the evening of 8-5-95 and lasting through noon the following day. After 1.2 cm of rainfall in the preceding two hours, discharge at the culvert increased from 6.8 l s^{-1} to 10 l s^{-1} (9:30 pm measurement, Table 2). In the following twelve hours, during which an additional 6 cm of rain fell, discharge increased from 10 l s^{-1} to $\sim 250 \text{ l s}^{-1}$. Twenty-four hours later, with little additional rainfall, discharge had declined to $\sim 28 \text{ l s}^{-1}$. This rapid and short-lived elevation of discharge probably reflects limited water storage capacity of thin rocky soils common in the catchment, saturation of lowland debris fan sediment, and possibly the large stream surface area associated with wide zones of shallow unchannelized flow in first-order tributaries. The concentration time calculated using Snyder's Method (Snyder, 1938) is 0.63 hours according to a 1989 engineering inspection report.

High suspended-sediment concentrations in streamwater are associated more directly with precipitation than with discharge. Sediment loads during February 1995 (discharge $\sim 30 \text{ l s}^{-1}$) were much lower than loads recorded during July 1994 (discharge $\sim$

TABLE 2. Precipitation, Discharge and Suspended Sediment Concentrations.

Date	Location	Time	Prec. (cm)	Disc. (L s ⁻¹)	Sed. Conc. (g L ⁻¹)	% Org.	Notes
7-20-94	culvert	12:00 pm	---	2.2	---	---	discharge only
7-21-94	culvert	5:00 pm	---	11.0	0.0541	29.7	t-storm
10-8-94	culvert	12:00 - 5:00 pm	none	1.6	0.0072	---	drought
2-18-95	culvert	morning	---	27.0	---	---	
2-20-95	culvert	morning	none	36.0	0.0048	---	
2-23-95	culvert	2:13 pm	none	34.5	0.0015	---	
6-18-95	culvert	3:20 pm	---	38.3	---	---	
6-21-95	culvert	11:45 am	none	20.3	---	---	
8-2-95	culvert	5:00 pm	---	5.7	---	---	
8-5-95	culvert	9:50 am	0.1	8.5	0.0116	21.5	overnight pptn.
	culvert	5:00 pm	0.2	6.8	---	---	1.5 hrs after t-storm
	culvert	9:30 pm	1.2	10.0	0.0464	22.2	2 hour pptn.

(TABLE 2 continued next page)

TABLE 2. (Continued)

Date	Location	Time	Prec. (cm)	Disc. (L s ⁻¹)	Sed. Conc. (g L ⁻¹)	% Org.	Notes
8-6-95	culvert	10:00 am	5.8-6.6	249*	0.0421	26.6	pptn. since 7:30 pm, 8-5-95
	above nick	12:15 pm	---	---	0.0296	29.7	
	below nick	12:20 pm	---	---	0.0313	29.7	
8-7-95	culvert	9:00 am	3.5-1.0	28.3	---	---	pptn. from 9:30 am to 3:00 pm on 8-6-95; water is clear
	lake surface	9:30 am	---	---	0.0111	21.6	
	near riser						
12-3-95	culvert	3:00 pm	---	40.0	---	---	
7-29-96	culvert	10:20 am	---	20.0	---	---	

* Discharge estimated using the Manning Equation.

11 l s⁻¹) and August 5th 1995 (discharge ~ 9 l s⁻¹) both associated with minor precipitation (Table 2). In addition, highest recorded sediment concentrations can be similar between different sizes of events. Suspended sediment concentration was nearly the same (~ 0.05 g l⁻¹) for storm-influenced discharges on 7-21-94 (11 l s⁻¹), 8-5-95 (10 l s⁻¹), and 8-6-95 (~250 l s⁻¹). Some of the variation is a result of sampling bias associated with a lack of continuous short-time-interval water sampling over the course of each precipitation event.

To get an idea of relative sediment contributions from first-order and rapidly incising portions of the basin, I sampled streamflow just upstream and downstream of the knickpoint during the August 5-6 storm (Table 2, Fig. 4). These knickpoint samples were similar in suspended load (~ 0.03 g l⁻¹), containing 75% of the suspended sediment content measured downstream at the culvert, and represent 38% of the basin area. (This basin area percentage considers only the south tributary which is the only surface water feed at this point in the catchment.) This apparent disproportionate representation of first-order streams as sources of suspended sediment downstream at the culvert does not account for sediment storage dynamics along the second-order channel or sediment composition. Percent organic matter of these samples was higher than that recorded at the culvert during this storm, suggesting that exposed streambanks along the second-order stream and the channel bed do act as important sources of suspended mineral sediment. Organic matter percentages for the knickpoint samples are roughly 8% higher than that of all other samples taken during August 1995, but compare well with the percentage obtained for the July 21, 1994 thunderstorm sample taken at the culvert, which had higher sediment concentration (0.05 g l⁻¹). Strong observational evidence for active erosion along the second-order channel relative to that along first-order channels suggests either the presence of a major upstream sediment source that is difficult to visually

observe, or that the two knickpoint samples could be unrepresentative (error associated with small sample size).

The day following the August 5-6 storm, I sampled lake surface water 10 m out from the impoundment to measure the amount of suspended sediment that could be leaving the lake water column through the outflow riser without being deposited. This single sample contained a total load of 0.0111 g l^{-1} , 21.6% of which was combustible organic matter. Making the simplistic assumption that no surface water was retained in the lake long enough for further settling of sediment and the depth of surface water lost with the above sediment concentration was 20 cm, the loss of sediment from outflow during the storm could be roughly 22% of input. This calculation assumes that suspended sediment measured at the culvert (0.05 g l^{-1}) was maintained during 8 hours of the 24-hour elevated discharge period, and that resuspended lakebed sediment represents a negligible contribution to water column sediment concentration. This computation also assumes that time elapsed was sufficient to permit dispersal of suspended sediment throughout the lake. This assumption is difficult to test; no clearly visible sediment plume was apparent anywhere outside the deltaic zone. Using these constraints, trap efficiency during moderate storm events would be 3% lower than that estimated for annual trapping using Heinemann's curve. For high-magnitude storms in which a greater proportion of total sediment yield is coarse-grained, trap efficiency might match or even exceed the Heineman curve value.

VI. FLOOD HISTORY RECONSTRUCTION

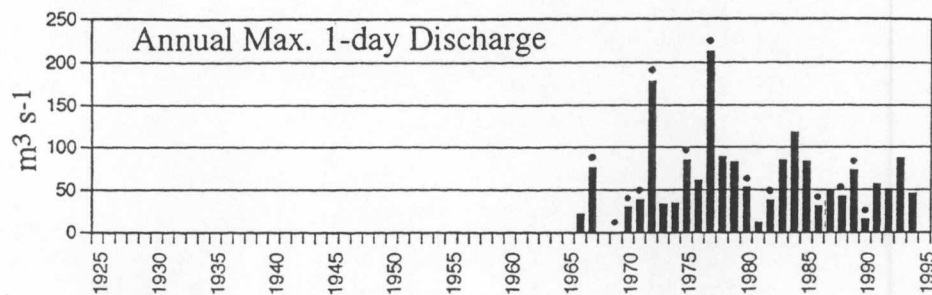
Precipitation and Discharge Data

Daily precipitation data from Mt. Weather, Virginia, provide the best available approximation to precipitation history at Thompson Lake. By comparing precipitation data from Mt. Weather to that from other nearby sites, and to discharge data from Goose Creek, which receives the drainage from this mountain region, regional rainfall events can be distinguished from isolated thunderstorms. Histograms of maximum one-day precipitation and coincident discharge, and maximum one-day discharge and coincident precipitation for each water year (October to September) are presented in Figure 5. This figure also shows a histogram of total precipitation by water year. Dots over bars indicate occurrence during the months of May through November, encompassing the growing season and period of low baseflow in Crooked Run. These precipitation and discharge characteristics were chosen as the ones with the greatest potential to be indicators of flood events in the Thompson Lake drainage basin. Other measures compiled but not displayed include maximum three-day precipitation by water year for all months in the water year, and for the growing and dormant seasons separately, and the number of days per water year in which one-day precipitation equaled or exceeded 1, 2, 3, 4, 5, 6 and 7 inches. The three-day measurements appeared to offer little additional information when compared with the one-day measurements, and the annual frequencies of various precipitation depths follow closely the record of total annual precipitation. More complex combinations of precipitation characteristics potentially indicative of flooding were not compiled, but could offer an improvement over the measurements used here.

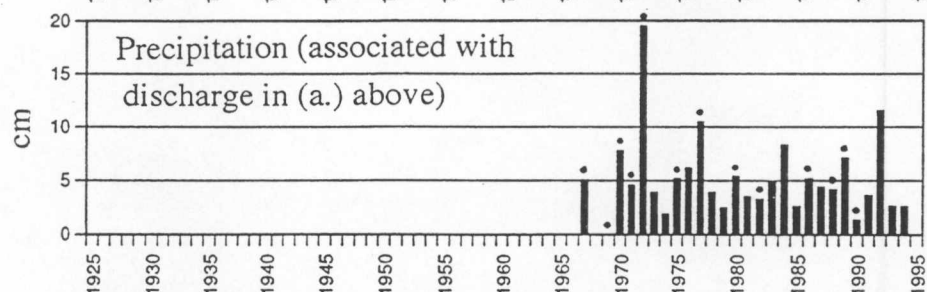
High annual maxima in one-day precipitation occurred primarily during the 1970s and in the early 1990s. During the 1980s, high one-day precipitation was relatively

FIGURE 5. Histograms of precipitation and discharge from the Mt. Weather and Goose Creek stations (respectively). Water year is October through September. Dot symbols over bars represent event occurrence during the months May through November.

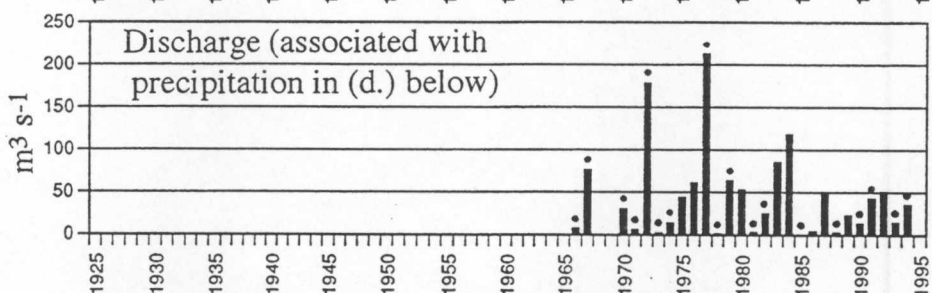
(a.)



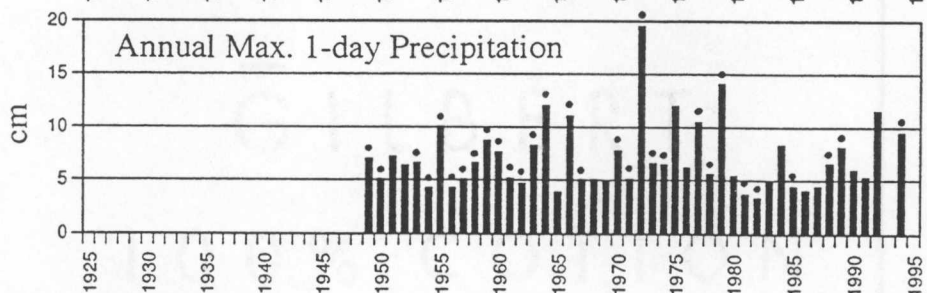
(b.)



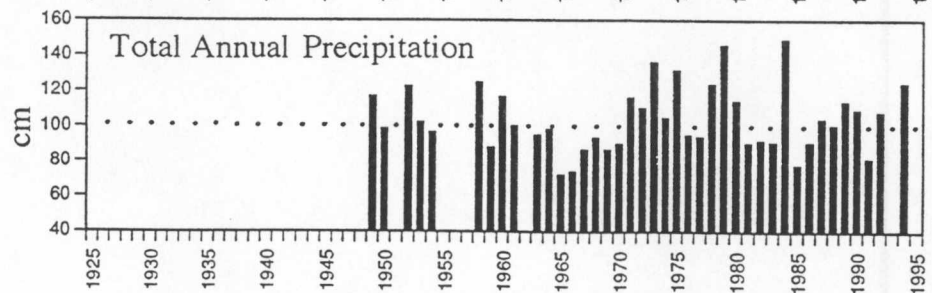
(c.)



(d.)



(e.)



Water Year

uncommon; 1984 and 1989 were exceptions, but are still represented by only moderately high values. Highest values in 1975, 1984 and 1992 occurred during the winter months and were therefore possibly part or all snowfall, potentially leading to reduced stormflow and lower peak discharge. This may be particularly true for the 1975 value which coincides with only moderate discharge in Goose Creek (Fig. 5). Sediment introduction associated with rainsplash on bare streambanks may also have been minimized under these conditions due to frozen ground cohesion. Alternatively, more frequent freeze-thaw cycles during this season may have loosened surficial soil, increasing its erosion (Wolman, 1959). Precipitation associated with Hurricane Agnes in July 1972 is clearly shown in Figure 5, as are summer storms in 1977 and 1979. Three-day totals coinciding with peak precipitation maxima are rarely much greater; most of the rainfall for annual peak maxima at this site falls during single-day events.

Highest values of the maximum daily average discharge at Goose Creek (Fig. 5) correspond to high precipitation for 1972 (Hurricane Agnes; summer), 1977 (autumn), and 1984 (winter). Values during the latter half of the 1970s are generally high ($\sim 80 \text{ m}^3 \text{ s}^{-1}$). Other relatively high discharge values occur for 1983 (winter), 1985 (winter), 1989 (summer), and 1993 (winter). The histogram of coincident precipitation for this discharge record (Fig. 5) indicates that winter peaks are not necessarily associated with large storms, and may mainly reflect high seasonal baseflow. The events for which high discharge and precipitation coincide are likely to be regional in extent and therefore suggest the occurrence of major discharge events at Thompson Lake. The most notable event years are 1972 (Hurricane Agnes), 1977 (summer), and 1984 (winter).

Although no discharge data are available for the Thompson Lake catchment, an idea of the flood magnitude associated with Hurricane Agnes for this size of basin can be obtained using the few existing records from other small catchments nearby. The south fork of Sycolin Creek near Leesburg, Virginia (5.3 km^2 ; 40 km northeast) had peak

discharge of $32 \text{ m}^3 \text{ s}^{-1}$ ($32,000 \text{ l s}^{-1}$) for precipitation similar to that at Thompson Lake ($\sim 20 \text{ cm}$). This value contrasts with plots of peak discharge vs. basin size for the Agnes event presented by Miller (1990) and Bailey et al. (1975). For catchments the size of Thompson Lake, peak discharge for Agnes intersects these two empirical curves at 70 and $110 \text{ m}^3 \text{ s}^{-1}$ respectively. If the Agnes event peak discharge at Thompson Lake was near the Sycolin Creek (nearest site) value of $32 \text{ m}^3 \text{ s}^{-1}$, the flood magnitude was over 100 times greater than that produced by the 6-cm storm I monitored in August 1995, even though the Agnes event precipitation was only three times greater. Differences in rainfall intensity, the larger size of the Sycolin Creek catchment and the dry antecedent conditions prior to the August 1995 event are at least in part responsible for this disproportion.

Dendrohydrology

For small catchments, like that of Thompson Lake, individual isolated thunderstorms may cause flooding without producing a large discharge peak as far downstream as Goose Creek. This means that even though the events inferred from precipitation and discharge data in the prior section likely indicate flood events at Thompson Lake, these events are not the only ones possible. Information from flood damaged trees along Crooked Run provides an additional check on the gauging data from Mt. Weather and Goose Creek, as well as information on local events. Because the number of damaged trees available for analysis is limited by the small size of Crooked Run, I also sampled trees from Wildcat Hollow, an adjacent and similar basin to the south (Fig. 2). Data from these two sites are presented in Figure 6.

FIGURE 6. Temporal distribution of tree scars and other types of flood damage to trees along channels of Crooked Run and Wildcat Hollow. (a.) Flood damage types differentiated for Crooked Run and Wildcat Hollow; (b.) Composite of dendrohydrological flood evidence for both basins.

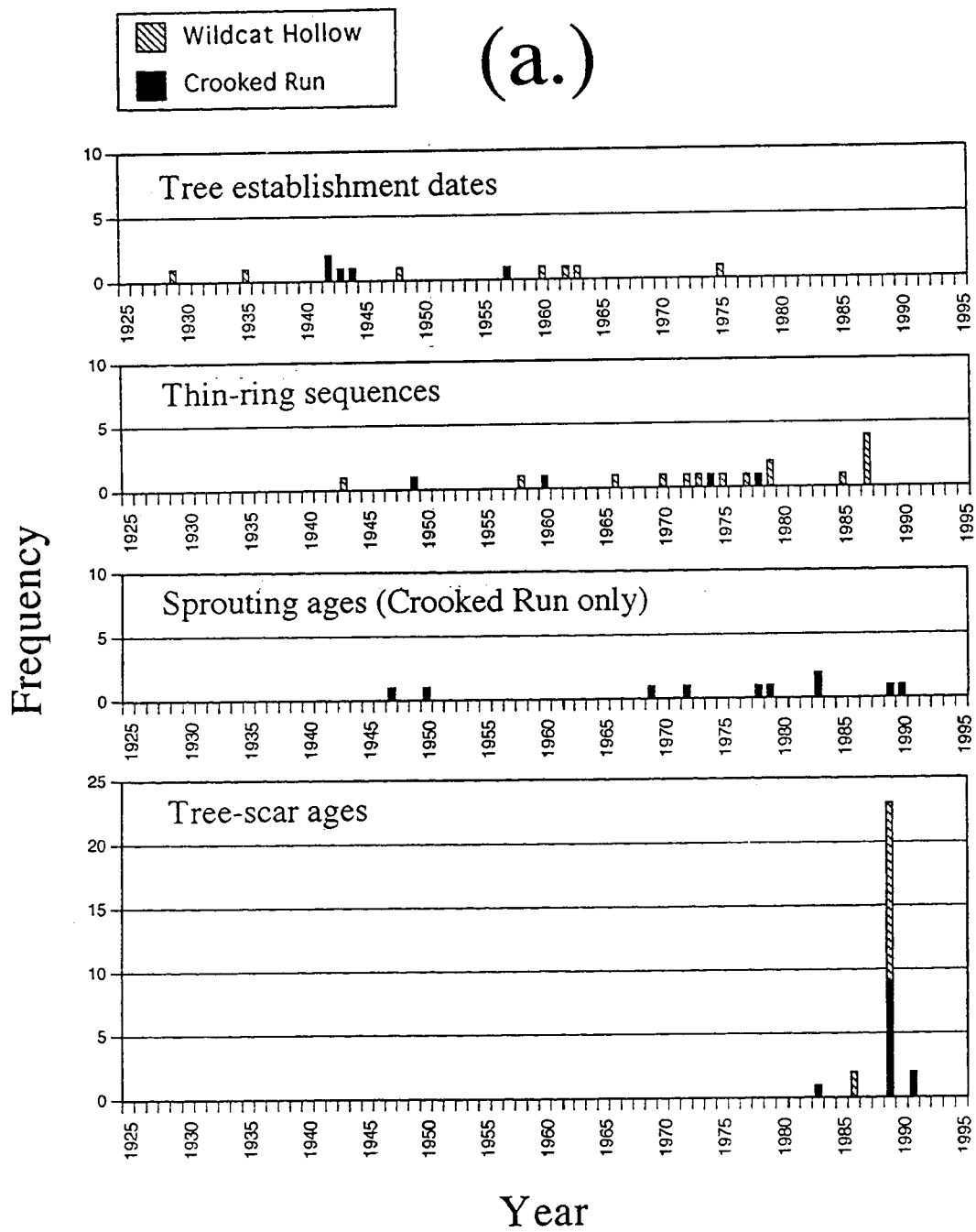
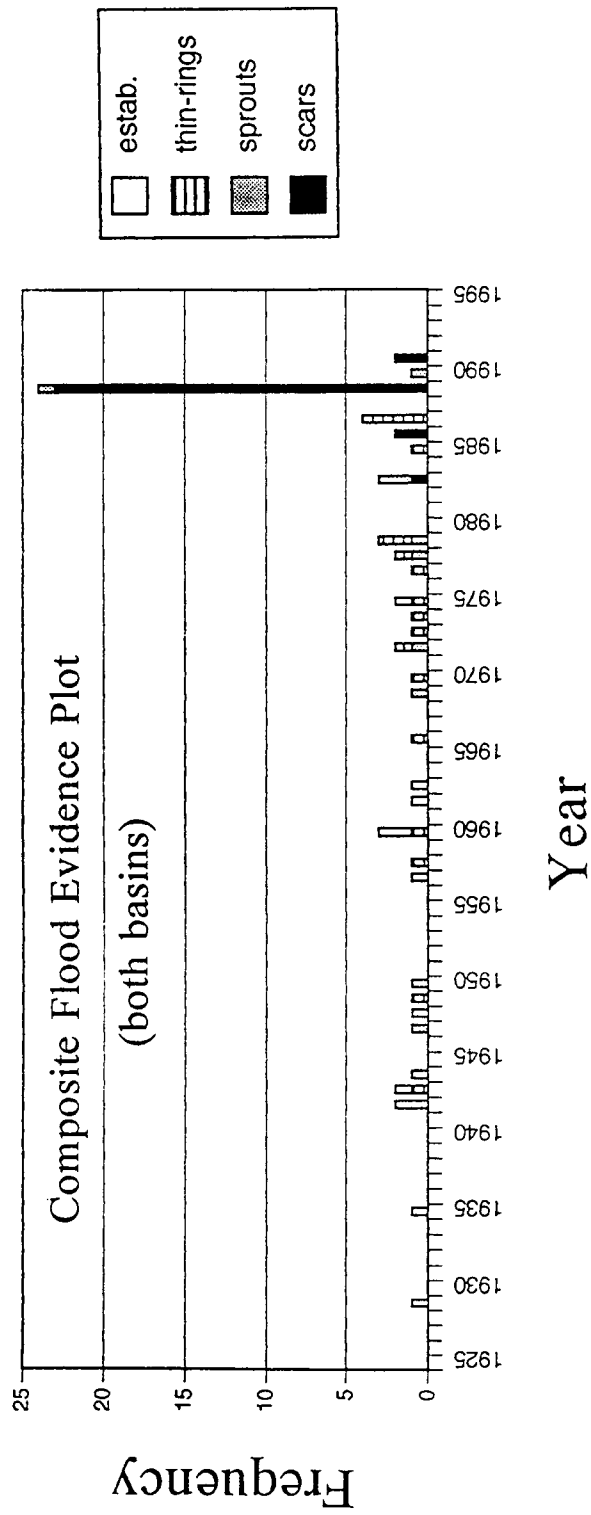


FIGURE 6 (continued)

(b.)



Several types of flood damage to trees may be visible along a single stream channel. Most notable are flood scars, which are markings or depressions devoid of bark and rimmed by a smooth rounded woody growth (callus) on the upstream-facing sides of tree boles. The date of creation of flood scars can be ascertained by tracing the tree-ring (the annual growth increment) containing the scar, in a wedge of wood taken from the scarred area, to an adjacent undamaged area, and counting the number of rings formed after the scar. Scars are formed by the killing of bark and underlying growth tissue (cambium) through the processes of floating object (e.g. woody debris and ice) impact and/or abrasion by suspended sediment. Scars formed by the former mechanism would be expected to mark the approximate flood height. Scars formed by particle abrasion usually form nearer the trunk base within the layer of saltating bedload. It is difficult to positively identify the mechanism responsible for creating a given scar. Damage to the tree bole that does not completely remove the bark or kill the cambium may, instead of scars, produce narrow rings that are also clearly visible in tree bole wedges and increment cores taken from the upstream sides of tree boles (Alestalo, 1971).

Flooding may also lead to bank erosion and subsequent collapse, causing trees along the channel walls to tilt or topple over into the flow. New limb sprouts, or sprouts from exposed root mass often become established within a couple of years, growing vertically from the inclined tree or exposed root. The number of tree-rings exposed in a wedge from such a sprout represents a minimum estimate of the number of years since tilting.

I obtained four types of data from flood damaged trees. The types are: 1.) age and estimated season of flood scarring, 2.) sprout ages from tilted trees along channel banks, 3.) ages of narrow ring sequences possibly indicative of severe bark damage, but short of complete removal, and 4.) tree age, representing a minimum time since either construction of, or complete vegetation removal from stream-side surfaces. Scar ages are

the highest quality data points because they directly date the occurrence of the associated scarring event. Sprout ages from tilted trees are of lesser quality because bank collapse following deep undercutting during a flood does not necessarily occur during the flood itself; also sprouting may not immediately follow tilting. The initiation dates of narrow ring sequences are well-established, but the reason for their formation cannot be uniquely identified. Finally, problems of interpreting establishment dates include sampling bias (explained below), the increased potential for error associated with longer ring counts, inability to crossdate, and failure to recover pith in every core. The inability to crossdate is a result of the variety of species and micro-habitats from which samples were derived, and the resulting variability in growth responses to similar stimuli.

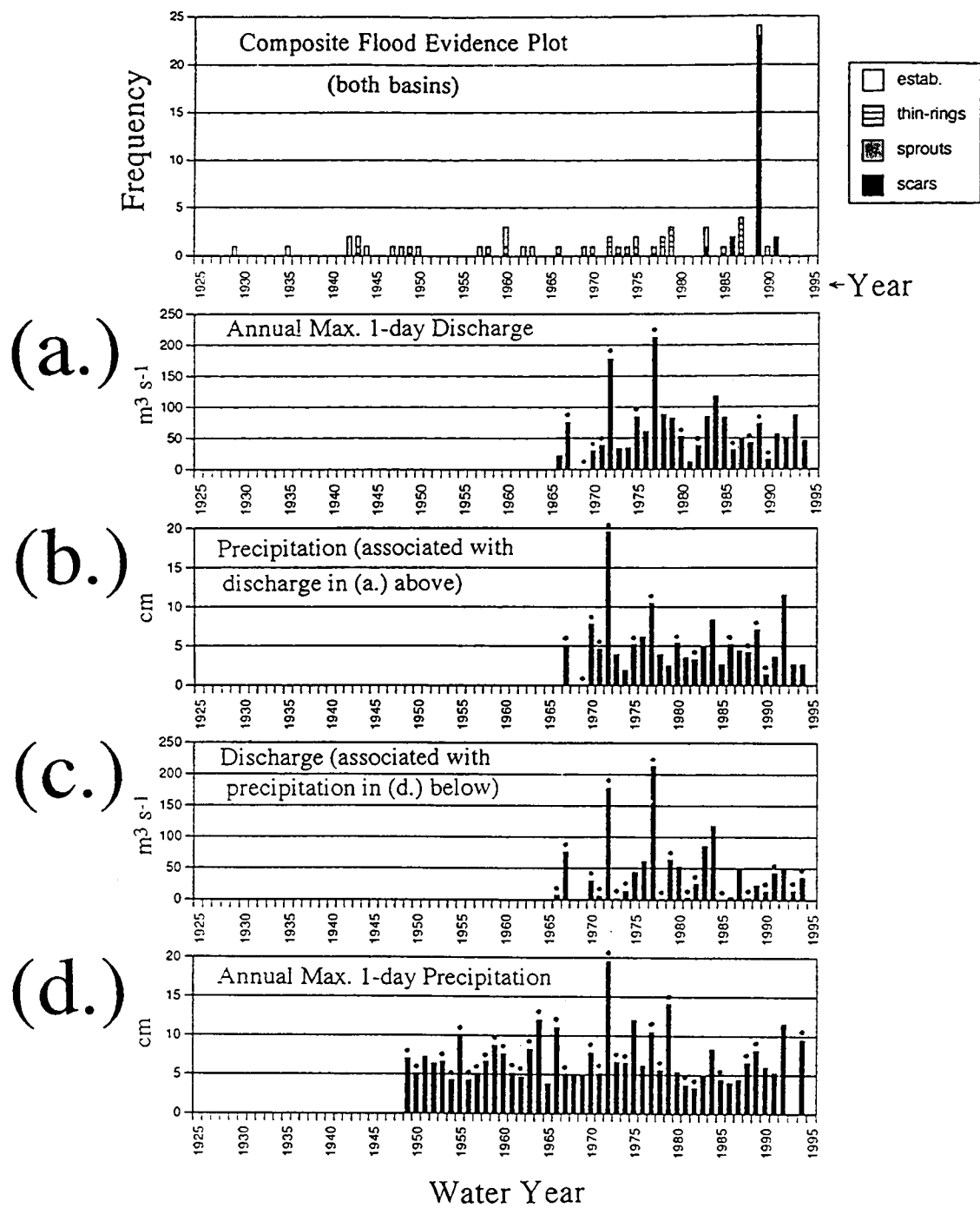
Tree scars at Crooked Run were formed between the 1983 and 1984 growing seasons, during the growing season of 1989, and between the 1991 and 1992 growing seasons; at Wildcat Hollow, scars date from between the 1986 and 1987 growing seasons and during the growing season of 1989 (Fig. 6). At both sites, scars dating to 1989 are by far the most common. Scars dating to 1986-87 and 1991-92 are fewer at two each, and there is only one scar dating to 1983-84. The dates from most sampled sprouts are earlier than 1984, although two sprout ages coincide with the scar date in 1983-84. I chose not to sample very young sprouts evident along recently collapsed channel walls; these sprouts appeared to be less than three years old. Initial dates of thin ring sequences and individual thin rings cluster at 1987 and are distributed singly throughout the 1970s. Tree establishment dates form no prominent clusters and are probably of limited use, especially considering the small number of ages obtained. Because I did not systematically sample multiple trees from a given landform, there is a sampling bias associated with these establishment ages; they reflect only those trees bearing scars.

Using the coincidence of scars, and scars with at least one other type of evidence as guidelines for selecting possible flood dates, four events are indicated: 1.) 1991-92

(two scars, both from Crooked Run), 2.) 1989 (23 scars total from both basins, one sprout), 3.) 1986-87 (two scars and four thin-ring sequences the following year, all from Wildcat Hollow) and 4.) 1983-84 (one scar and two sprouts, all from Crooked Run). Note that the only event year for which there is evidence in both basins is 1989. Because sample numbers are small for the other possible dates, it is difficult to tell if this mutual exclusion represents sampling error, different flood histories, or scar generation unrelated to flooding.

In Figure 7, I compare these possible flood dates to the precipitation and discharge records of Figure 5. A dot at the top of a bar represents occurrence in the months of May through November, roughly coinciding with the growing season. Corresponding maximum one-day precipitation totals occur for all hypothesized event years except 1986-87, if season of scarring is accounted for (Fig. 7). That is, for 1991-92 scars which formed after the 1991 growing season but before observable growth in 1992, there is a corresponding (and potentially related) high-precipitation event in winter (however, note the absence of coincident high discharge at Goose Creek during this period; Fig. 5). The same situation occurs for the 1983 scar event. The clarity and abundance of evidence for the 1989 event makes its interpretation disproportionately important to the understanding of hydrological processes and environmental history. This year is unusual in that given the large numbers of scars formed, it might be expected that a larger than observed precipitation peak would occur for this year (remembering that 1989 scars were formed during the growing season). Maximum one-day precipitation is only moderate for 1989, and coincident Goose Creek discharge is low. One possible explanation for this scarring event is that in the 15 days following the 1989 precipitation maximum (8.1 cm on May 2), 19.8 cm additional rain fell during three storms at Mt. Weather. If similar or greater quantities fell at Crooked Run, sustained high flow resulting from constantly high antecedent moisture levels might explain the number of

FIGURE 7. Dendrohydrological evidence (from Fig. 6) compared with precipitation and discharge records from Mt. Weather and Goose Creek (from Fig. 5).



scars found. Another alternative is suggested by the record of mean monthly discharge for Goose Creek (not reproduced here). This record (along with many other discharge records from Virginia) distinguishes the year of 1989 as one of sustained discharge for the latter half of the water year, as much as 400% above average (Water Resources Data, USGS, 1990). During most other years, peak flows correspond to high precipitation onto saturated ground in late spring. The year of 1989 contrasts with these years by having higher discharge than usual later in the growing season. Abnormally high precipitation (29.5 cm) during May at Mt. Weather was followed by 11.4 cm in June and 17.8 cm in July. Antecedent moisture and summer baseflow may have been sufficiently high during this time to produce substantially elevated discharge during even minor storms. Persistent wet conditions late in the growing season are consistent with the season of scarring (mid- to late- growing season, 1989) indicated by the position of a scar within a year's growth increment (Alestalo, 1971).

The lack of a coincident precipitation peak for the 1986-87 scar event that is evident only in Wildcat Hollow, could indicate the occurrence of a locally intense thunderstorm not sampled at Mt. Weather (14.5 km away). This storm would have had to be local because widespread rainfall would likely have produced a higher peak discharge at Goose Creek than observed. Another possibility remaining involves the elevation of scars above the channel; low scars could have formed under lower flow conditions. However, scars from 1986-87 are high, at 60 cm and 120 cm above the current channel. Such flood heights in the unconstricted channel area where the scars occur suggest high discharges not consistent with the peak rainfall observed at Mt. Weather during the time of scar formation (between 1986 and 1987 growing seasons). A plausible alternative is that the scars did not originate by flooding, but instead through potential disturbances from insect pathogens or animal activities. The small number of scars representative of this event hinders the interpretation of their cause.

The Mt. Weather data also record several peaks in maximum annual one-day precipitation during the 1970s, including nearly 20 cm of rainfall associated with Hurricane Agnes in 1972. No tree scars date to the 1970s, although many thin-ring sequences and a few sprouts originated during this decade. The lack of tree scars dating to earlier than 1983 is notable, especially considering the size of the Hurricane Agnes precipitation and discharge peaks in 1972. Sigafos (1964) found similarly youthful scar ages in three small Maryland basins. It is more likely that scars originating in 1972 have healed over, so that I missed them in my sampling. This is consistent with healing rates I have observed on the 1989 scars. For the small trees typical of stream marginal sites in this area, approximately half the scar areas from 1989 scars are healed over six years later. Although it might be expected that healed 1972 scars would be visible in increment cores through scars formed in the last decade, 1972 scars would have had to form at nearly the same bole height as the more recent scars to be directly observed. It could be that the thin rings prominent in the 1970s represent zones of minor damage near the elevation of more recent scarring.

Elevation of scars above the channel reflects the minimum depth of flood waters at a given channel cross section. Complicating this simple idea, tree scars from flooding can be produced by multiple mechanisms (Alestalo, 1971). For example, if scars are produced by impacts with floating debris, they should reflect full water depth. In contrast, scars produced by corrosion from traction or saltating bedload form at some fraction of the full water depth (Alestalo, 1971). Furthermore, a scar found near the base of a bole growing near the current channel level could have formed through bedload corrosion during a large flood, or through floating object impact during a smaller flood. Scars beneath woody debris piles in the Crooked Run and Wildcat Hollow basins suggest that in some cases, impact and corrosion from snagged woody debris might be

responsible for scarring. Pile-up of such debris above water line could result in scars being elevated slightly above the floodwater surface.

I calculated average scar elevations above channel for each potential flood year except 1983 (only one date) to compare with annual maximum daily precipitation totals at Mt. Weather. For the 1986-87 event, I included scar elevations at which year-1987 thin rings, which might be indicative of widespread minor damage to the cambium, were found. As might be expected, considering the small number of samples for event years other than 1989, plots of average scar elevation vs. precipitation maxima (not shown) have high scatter. Scar elevations indicate water depths substantially greater than the largest flow event I have monitored (~ 30 cm average depth following a 7-cm, twelve-hour storm, August 1995) at Crooked Run. For the 1989 flow event, the average elevation for scars from both Crooked Run and Wildcat Hollow indicates average minimum flow depths roughly double those I measured for the August 1995 storm. The 1989 precipitation maximum (8.1 cm on May 2) is comparable to that of the 1995 storm (seven centimeters), suggesting that this (1989) event is not likely to be the single cause for the 1989 scarring. I therefore regard extended high summer precipitation (described earlier for this year) as a more probable explanation for this vegetation damage.

In addition to sampling biases mentioned earlier, several limitations inhibit the interpretation of the dendrohydrological data presented in Figure 7. The main problems are associated with temporal and spatial resolution of rainfall data and the associated ability to provide useful measures of effective precipitation (contributing to stormflow). Similar rainfall amounts may be produced by slow steady rain all day, or one 30 minute high-intensity storm; the runoff response would differ in each case. I obtained 15-minute precipitation data from Mt. Weather to analyze storm intensity, but the data are patchy and do not cover any of the four hypothesized flood dates. A major factor in the ability to compare rainfall records to tree disturbance data is the distance from the basin to the

weather station. In this study there is evidence for a flood event during the 1989 growing season, even though the maximum daily precipitation at Mt. Weather during this year is not high. The same situation exists for 1986-87 scars, although there is less tree-disturbance evidence for this event.

In summary, there is tree-ring evidence for flooding in 1991-92, strong evidence for a major event during the 1989 growing season, and weaker evidence for floods in 1986-87, and 1983-84. Small numbers of scarred trees and rapid scar healing on the small trees typical of the stream margins at this site are obstacles to flood history reconstruction. It seems logical that the number of scarred trees along a channel (per unit length) would be influenced by channel morphology, flood frequency and intensity, and the rate of scar healing. The time required for a scar to heal over is related to the size of scar, which is limited by tree diameter and intensity of corrosion. Age and growth rate of a species might also influence scar healing time. Alternatively, floodplain species may respond similarly to corrosion with regard to allocation of resources to callus formation. If small trees predominate along stream marginal sites for ecological reasons, maximum scar size and therefore scar longevity may be intrinsically limited. The lack of old scars might be corrected by careful sampling of stream-side trees not currently bearing visible scars.

With regard to interpreting the sediment yield record, the discovery of the 1989 event, not obviously indicated by off-site precipitation and discharge records, but for which there is abundant dendrohydrological evidence, demonstrates the potential utility of dendrohydrologic techniques, even for small streams. In the absence of other means to assess the validity of using off-site monitoring records from Mt. Weather and Goose Creek to understand flood history at Thompson Lake, I will proceed according to the assumption that these records represent reasonable approximations. In the next chapter, I lay out the sediment yield reconstruction that will later be compared to the flood history.

VII. LAKE SEDIMENT ANALYSIS

Sources of Sediment in Thompson Lake

Sediment deposited in Thompson Lake is potentially derived from four main sources: lake bank erosion, airborne fallout, biogenic silica in the water column, and erosion of the upstream catchment. Autogenic secondary mineral formation appears minimal within sediment deposits (see magnetic susceptibility section) and does not represent a separate input. Soil erosion from the lake banks is primarily a result of wave action; its contribution can be estimated from erosion pin data (Table 1). Because only one year of erosion pin data is available from this study, this data source can now only serve as an indicator of the presence or absence of rapid bank retreat. The small amount of retreat suggested by the erosion pin measurements in Table 1 indicates that lake banks are eroding only slowly. However, even slow gradual retreat may substantially add to lake sediment. Dearing and Foster (1986) constructed a nomogram displaying the importance of the lake bank erosion input relative to catchment sediment yield, based on lake geometry, lake area to catchment area ratio and an estimate of sediment yield from the lake or other nearby monitored catchments. This nomogram was constructed assuming 40-cm-high erodible lake banks and a dry soil bulk density of 2.65 g cm^{-3} , however, the nomogram can be adapted to other conditions. The key element of this nomogram is the calculation of sediment mass introduced by lake bank erosion using the length of erodible bank (lake perimeter), height of bank, and the erosion increment (bank retreat). The current Thompson Lake perimeter (inclusive of the island perimeter) is 767 m, and the average bank height is estimated at 25 cm (although this has not been carefully surveyed). The limited erosion pin data indicate an estimated bank retreat rate of 0.84 cm yr^{-1} (0.0084 m yr^{-1}). Multiplying bank length times bank height and recession rate gives a

lake bank volume input of 1.61 m^3 ; multiplying this value times the dry bulk density 1.8 g cm^{-3} (equivalent to 1.8 t m^{-3} ; an estimate for silicate mineral soils at Thompson Lake) gives a mass input rate of 2.9 t yr^{-1} for lake bank sediment yield. This value for lake bank sediment yield is representative of current conditions at the lake, and may not apply for different site conditions, such as those that existed directly following reservoir construction.

Airborne mineral sediment inputs estimated from one year of direct (time integrated) sampling are low ($1.03 \text{ g m}^{-2} \text{ yr}^{-1}$). This value represents an input onto the lake surface of 0.0332 t yr^{-1} . The single sampling bucket used in this study and the short observation time reduce the reliability of this estimate. The measured value is, however, consistent with expected low fallout in humid forested regions.

Diatom frustules composed of biogenic silica do not appear to be an important source of lake sediment at Thompson Lake. The abundant diatom frustules that I have noted in loss-on-ignition residues (after 1000°C) from organic-rich lake sediments in small New York lakes (see studies by Clark et al., 1996; Clark and Royall, 1995) are not apparent in residues after 550°C from this site. The biogenic silica component of sediment input is therefore assumed to be negligible.

General Attributes of Thompson Lake Sediment Cores

A total of 33 sediment cores, including five supplemental cores from the three master core locations, were retrieved from Thompson Lake over a 22 month period (October, 1994 to July, 1996). Twenty-five of these cores were retrieved during two coring trips in the latter half of 1996. Cores vary in length from 7 to 35 cm and in all cases are composed of loose dark upper lake sediment, and basal dense cohesive soil representative of the

original (predepositional) lake bed. Length of retrieved cores is smallest near shorelines where bottom erosion is most pronounced and water column thickness low, and greatest in the deltaic zone. Almost all lake sediment cores are visibly homogeneous; exceptions are deep water cores A2 and A3 (locations shown in Fig. 3), which exhibit some unexplained centimeter-scale layering in upper portions. Cores from the delta region are distinguished by their visibly larger particle sizes of both mineral and organic debris fractions.

Magnetic Susceptibility

Down-core variation in magnetic susceptibility is shown for all cores at 1-cm intervals in Figure 8. Figure 8a shows profiles of mass specific low frequency magnetic susceptibility (X^{lf} ; referred to more simply as specific susceptibility) and Figure 8b shows profiles of frequency dependence of susceptibility (X^{fd}). The left ends of horizontal (susceptibility) axes on the profiles mark the approximate relative positions of coring sites. The controls of magnetic susceptibility in order of decreasing importance are the concentration of magnetic minerals, the size of magnetic particles, magnetic mineral type, and the shape of magnetic grains (Thompson and Oldfield, 1986; Dearing, 1994). All of these attributes are influenced by the abundance and type of magnetic minerals in the parent material of eroding soils, and weathering processes within the soil column leading to horizonation and the formation of secondary magnetic minerals. In addition, magnetic mineral concentration in lake sediment is affected, through dilution, by changing input of both allochthonous and autochthonous organic matter.

FIGURE 8. Magnetic susceptibility profiles for lake sediment cores. (a) Mass specific low frequency magnetic susceptibility (X^{lf}); (b.) Frequency dependent susceptibility (X^{fd}). Locations of axial vertices on each diagram approximately correspond to the spatial arrangement of coring sites shown on Figure 3.

(a.)

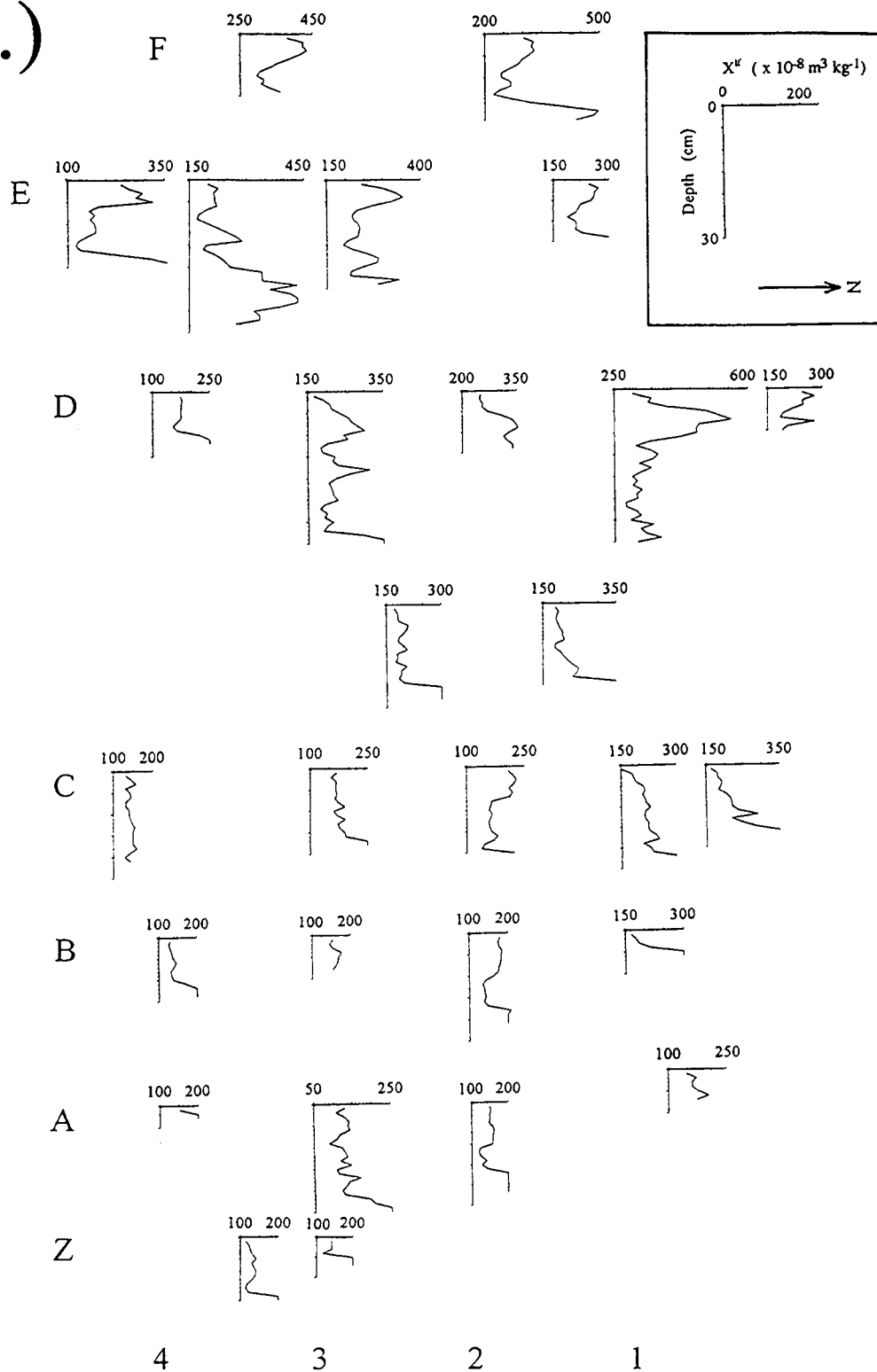
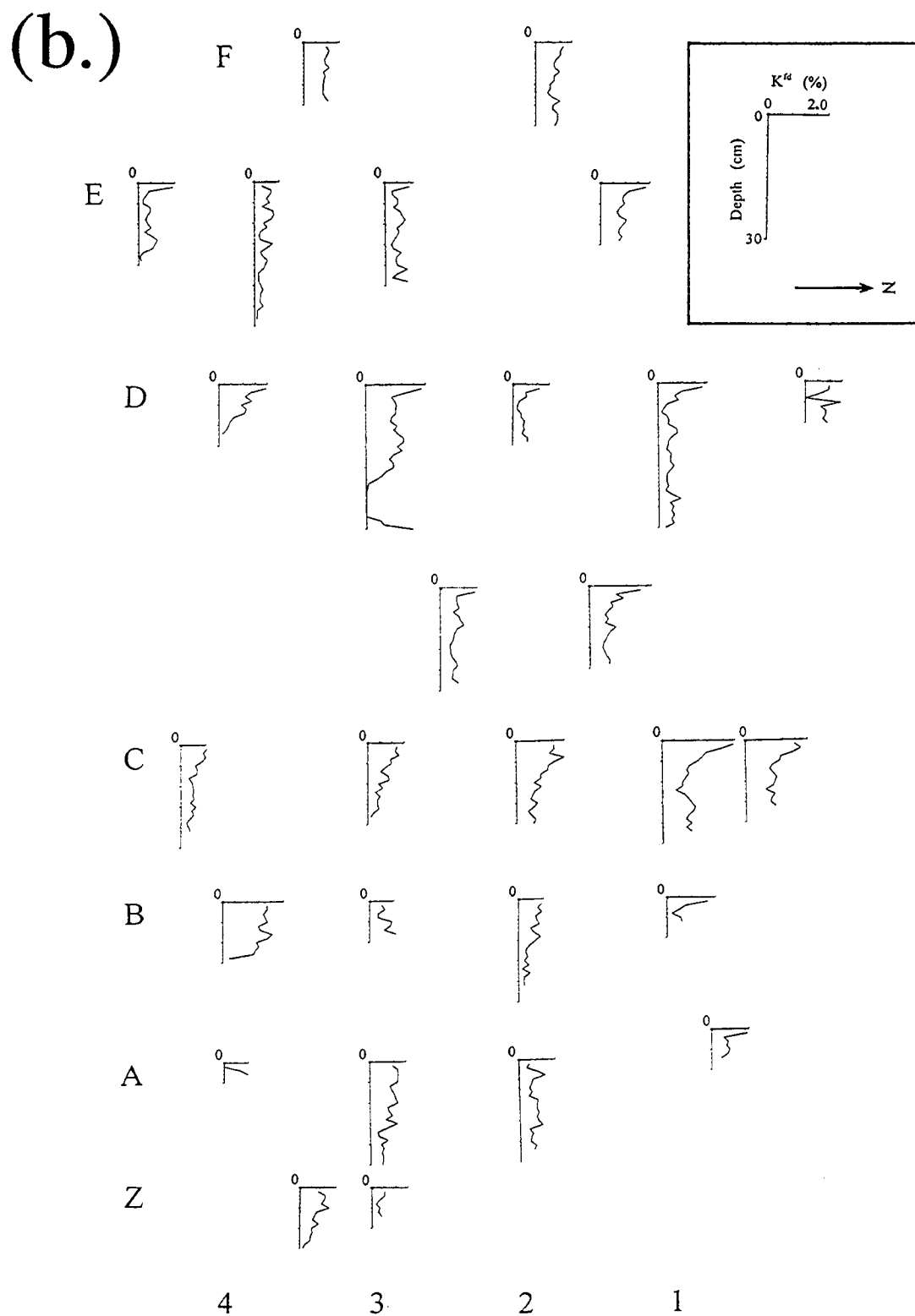


FIGURE 8 (continued)



The size of magnetic grains is of particular importance for lake sediment studies because it is in part controlled by the flow velocity of streamflow into the lake, and should increase during floods. Although the relationship between magnetic grain size and susceptibility is generally known, interpreting the susceptibility of natural samples is complicated because magnetic grains in most rock types are much smaller than the size of mineral particles released by weathering and transported by streams. At Thompson Lake, maximum and usually also minimum values within cores increase from deep water near the dam toward the lake inlet where flow velocity is greatest (Fig. 8a). This trend suggests that X^{lf} increases with particle size at this site. Reinforcing this suggestion are the trends in frequency dependent susceptibility (X^{fd}), which are controlled by the abundance of very fine superparamagnetic grains often of secondary origin. Figure 8b shows that values of frequency dependent susceptibility in all cores are low (less than 2%) reflecting the dominance of large primary magnetic minerals over fine secondary minerals formed as weathering products in soils and sediment (Dearing, 1994). In addition, average X^{fd} values are lowest in the coarse sediments of the deltaic area, and for individual cores, are most often in opposition to down-core variations in specific susceptibility (compare Figs. 8a and 8b). Lower values for X^{fd} toward the lake inlet indicate the apparent loss of fine magnetic grains to competent flow in this area, and the increasing dominance of coarse particles. Dearing et al. (1985) have shown similar trends at various European sites, and Dearing and Flower (1982) document such a relationship for silts from basaltic soils. The source material for these silts could have mineralogical similarities to greenstone soils at Thompson Lake.

Further support for the increase in X^{lf} with particle size at Thompson Lake comes from the characteristics of nine soil and channel sediment samples from the catchment. Samples from the clay-rich (field texture) C-horizon in the valley train deposit along the second-order channel had average X^{lf} of $63 \times 10^{-8} \text{ m}^3 \text{ kg}^{-1}$. Coarser A and B horizons

had average values of 310 and 233 ($\times 10^{-8} \text{ m}^3 \text{ kg}^{-1}$) respectively, higher than values for the C-horizon. Channel sediment samples deficient in fine-grained sediments as a result of fluvial sorting had the highest average X^{lf} of 484 ($\times 10^{-8} \text{ m}^3 \text{ kg}^{-1}$). Frequency dependent susceptibility (X^{fd}) values were low (maximum value 2.23%) for all sediment sources, indicating that secondary magnetic minerals are unimportant in catchment soils, and explaining the low X^{fd} values typical of all lake sediment cores. This lack of source differentiation would make the future application of magnetic mixing models for defining source areas problematic.

This circumstantial evidence for particle-size relations is supported by direct measurement of grain size distribution in lake sediment of known susceptibility. I conducted particle-size analyses for both high- and low-susceptibility zones in both deep water and deltaic zone areas (Fig. 9). Low- and high-susceptibility samples in cores A3 and B2 (deep water), D1 and E3.8 (deltaic) have mean grain sizes of 5.7 and 5.7 phi (A3), 5.8 and 5.7 phi (B2), 5.6 and 4.3 phi (D1) and 5.6 and 4.8 phi (E3.8). Magnetic susceptibility in almost all cores varies directly with mean particle size; core A3 appears to be an exception. Closer inspection of particle size distributions from core A3 (Table 3) reveals clear differences in particle size between the low- and high-susceptibility samples that do not manifest as differences in graphic mean. Sediment from 16-17 cm depth in core A3 ($X^{lf} = 160 \times 10^{-8} \text{ m}^3 \text{ kg}^{-1}$) contains higher total sand and less silt and clay than sediment from 14-16 cm ($X^{lf} = 120 \times 10^{-8} \text{ m}^3 \text{ kg}^{-1}$). These patterns suggest that deep-water X^{lf} is sensitive to input of sand-sized sediment and is usually lowered by abundant clay (clay percentage shown in Table 3). The abundance of sand varies directly with X^{lf} not only in deep-water, but also in all deltaic samples (Fig. 9).

Linear correlation coefficients computed for X^{lf} vs. weight percent of particulate size fractions in core samples corroborate the inference of high susceptibility for coarse

FIGURE 9. Mean particle size and weight percent sand fraction in relation to changes in X^{lf} for deep water (A3, B2) and deltaic (E3.8, D1) cores. First number given is mean particle size in phi units ($-\log_2$ of particle diameter in mm); second number is weight percent sand. An asterisk by a number pair denotes a sample that may have lost some fine particles through peroxide boil over during the removal of organic matter in sample pretreatment.

$X_{lf} \text{ (x } 10^{-8} \text{ m}^3 \text{ kg}^{-1}\text{)}$

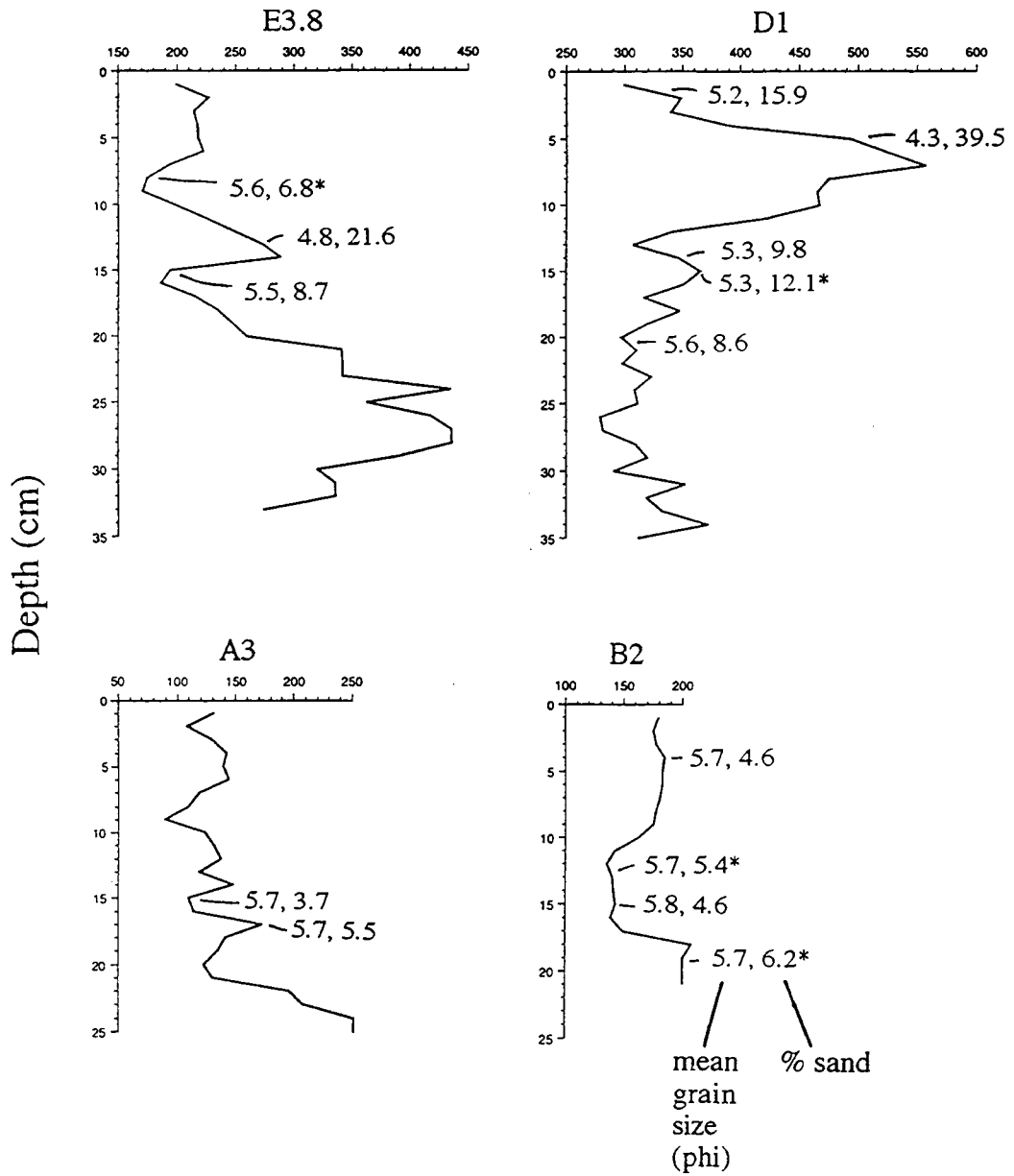


TABLE 3. Particle-Size Percentages for Selected Depth Intervals in Thompson Lake Cores.

SAMPLE	SAND				SILT		CLAY total
	vcos	cos	ms	fs	csilt	fsilt	
B2 3-5 cm	1.1	0.1	0.3	0.7	16.8	56.0	22.7
B2 12-14 cm	0.0	0.1	0.1	1.1	16.4	52.8	25.4
B2 14-16 cm	0.2	0.3	0.1	0.7	14.6	56.1	24.9
B2 18-21 cm	0.1	0.0	0.0	1.4	16.2	50.9	26.7
D1 0-3 cm	0.5	0.9	0.6	3.9	22.0	42.1	20.3
D1 5-7 cm	2.8	3.5	2.3	13.5	18.2	26.1	16.2
D1 13-14 cm	2.6	0.3	0.1	1.0	19.2	50.7	20.4
D1 14-16 cm	0.5	0.4	0.4	2.0	22.3	44.2	21.4
D1 20-22 cm	1.0	0.6	0.3	1.1	18.4	51.2	22.1
A3 16-19 cm	0.2	0.2	0.1	1.2	15.9	55.2	23.5
A3 14-16 cm	0.0	0.1	0.1	1.0	17.3	56.3	26.5
E3.8 7-9 cm	0.2	0.5	0.3	2.1	16.5	53.7	23.0
E3.8 12-13 cm	10.2	1.7	0.7	3.2	16.0	42.2	20.4
E3.8 14-16 cm	3.0	0.7	3.2	1.4	17.3	51.6	22.8

Abbreviations: vcos - very coarse, cos - coarse, ms - medium sand, fs - fine sand, vfs - very fine sand, csilt - coarse silt, fsilt - fine silt.

samples. Coefficients for sand fractions are positive and high (> 0.770) except for medium (0.381) and very coarse (0.307) sands. Highest correlation occurs for very fine sand (0.929). High correlation also occurs for mean grain size (-0.864 ; negative coefficient results from use of phi scale for mean grain size; see page ix, list of symbols and abbreviations). Correlations are positive and moderate in the relatively minor ($<20\%$) coarse silt fraction. Correlations for total weight percent of silt (dominated by fine silt, by weight), fine silt, and clay are all high (~ -0.850) and negative, demonstrating the importance of sand-sized particles for high susceptibility. A similar relationship was given by Björck et al. (1980) who found highest susceptibility values to be associated with coarse silt and fine sand fractions differentiated from glacial and fluvial sediments derived from igneous parent material.

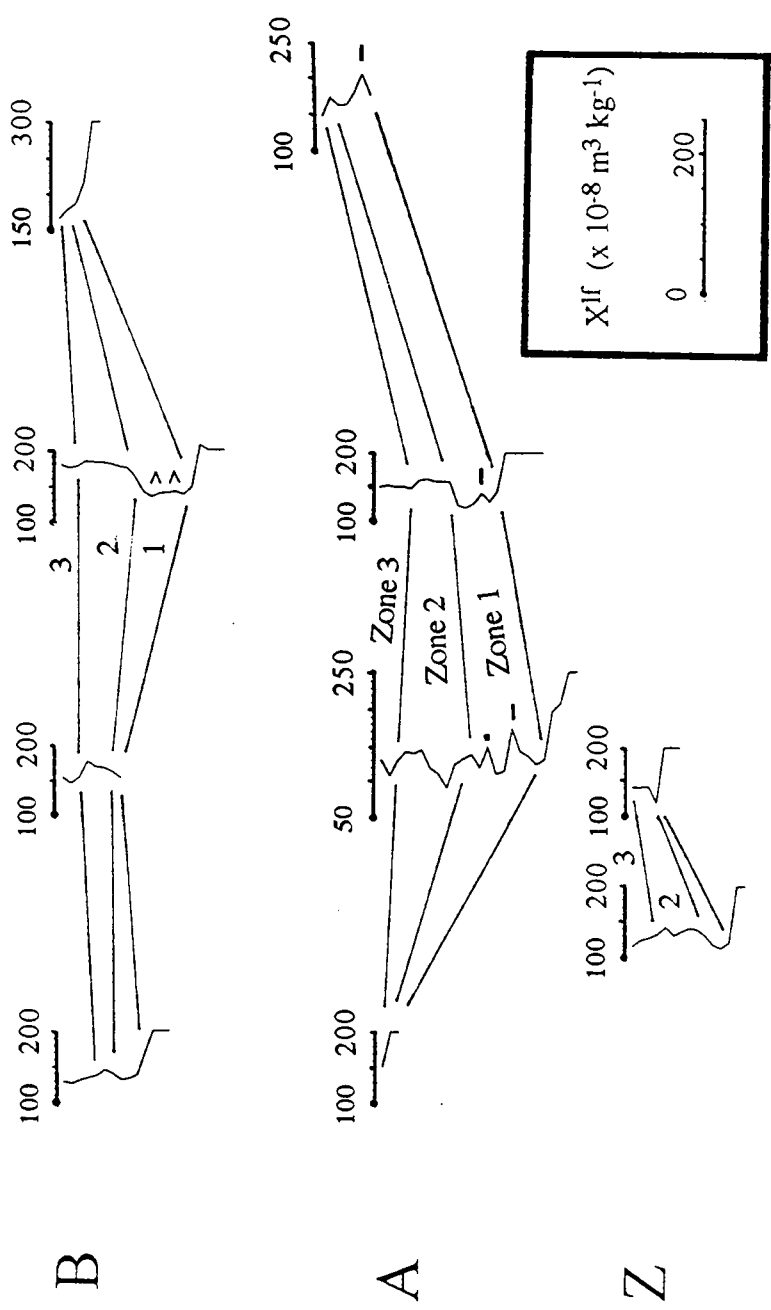
These analyses demonstrate that throughout the lake, high-susceptibility zones are associated with coarser sediment. Based on spatial trends in mass-specific and frequency-dependent susceptibility at Thompson Lake, and on laboratory analyses of lake core strata, particle size is the dominant control of magnetic susceptibility fluctuation in sediment profiles at this site. Dearing et al. (1985) state that this is typical for catchment soils for which the concentration of secondary magnetic minerals is low relative to that of primary minerals. Frequency dependent susceptibility measurements from seven catchment soil samples including hillslope and streambank topsoil and streambank subsoil and saprolite exposed to fluvial erosion are all less than 2%. These low values indicate that secondary soil minerals contribute little to specific susceptibility of catchment soils at Thompson Lake. I interpret fluctuations in specific susceptibility within cores to reflect changing particle sizes introduced into the lake by streamflows of differing competence.

Lake Sediment Stratigraphy

Down-core fluctuations in susceptibility are the basis for assigning magnetostratigraphic zones traceable between cores throughout the lake (Fig. 10). Two separate subgroups of profile shape can be distinguished, however, associated with proximity to the delta region. The first subgroup is associated with deep-water cores of transects Z, A, B, and C, (Fig. 10). These profiles do not exceed 20 cm thickness and I separated each into three major zones by qualitative inspection (see chapter 4 for further explanation). The basal zone (zone 1) lies directly above the sediment-soil interface, which is marked by an abrupt change in susceptibility, bulk density, organic matter, and color. This zone is comprised of a sediment layer of low background susceptibility, often with two positive fluctuations. These fluctuations are marked using dashes, dots and "V"s on Figure 10, and are hereafter referred to as events S1 (earliest) and S2 (most recent). Zone 2 is characterized by rapid increase from low zone 1 background values to sustained high susceptibility. These high values are sometimes interrupted by a minor decrease early in the zone, dividing zone 2 into two closely-spaced high susceptibility intervals. Zone 3 is the uppermost zone characterized by moderate susceptibility or, simply, declining values from zone 2 to the sediment-water interface.

The second subgroup of profiles is associated with cores from the delta region (transects D, E and F) which are characterized by two usually broad peaks near the surface, following an earlier period of low values (Fig. 10). Despite differences in shape attributes with respect to the first subgroup (from the deep-water area), the general simplicity of all profiles allows the two subgroups to be correlated. Zone 1 in the deltaic cores has low susceptibility relative to zone 2, but generally has absolute values higher than zone 1 in the deep-water cores. Two susceptibility peaks within deltaic zone 1 are similar to those in the lower sediments (zone 1) of the deep-water area; locations of these

FIGURE 10. Core correlation. Each sediment / stratum is numbered. Dashed lines in transect E marks correlation "through" the island. Two susceptibility fluctuation "events" are indicated by dots and dash marks beside the depth increments over which the events are sustained (i.e., the number of dashes corresponds to the thickness in centimeters of the event stratum). Dashes indicate the earliest event (event S1) and dots mark a later event (event S2). Small horizontal "V"s mark locations for which a thickness of 0.5 cm was used in the calculation of event sediment volume.



4 3 2 1

FIGURE 10 (continued)

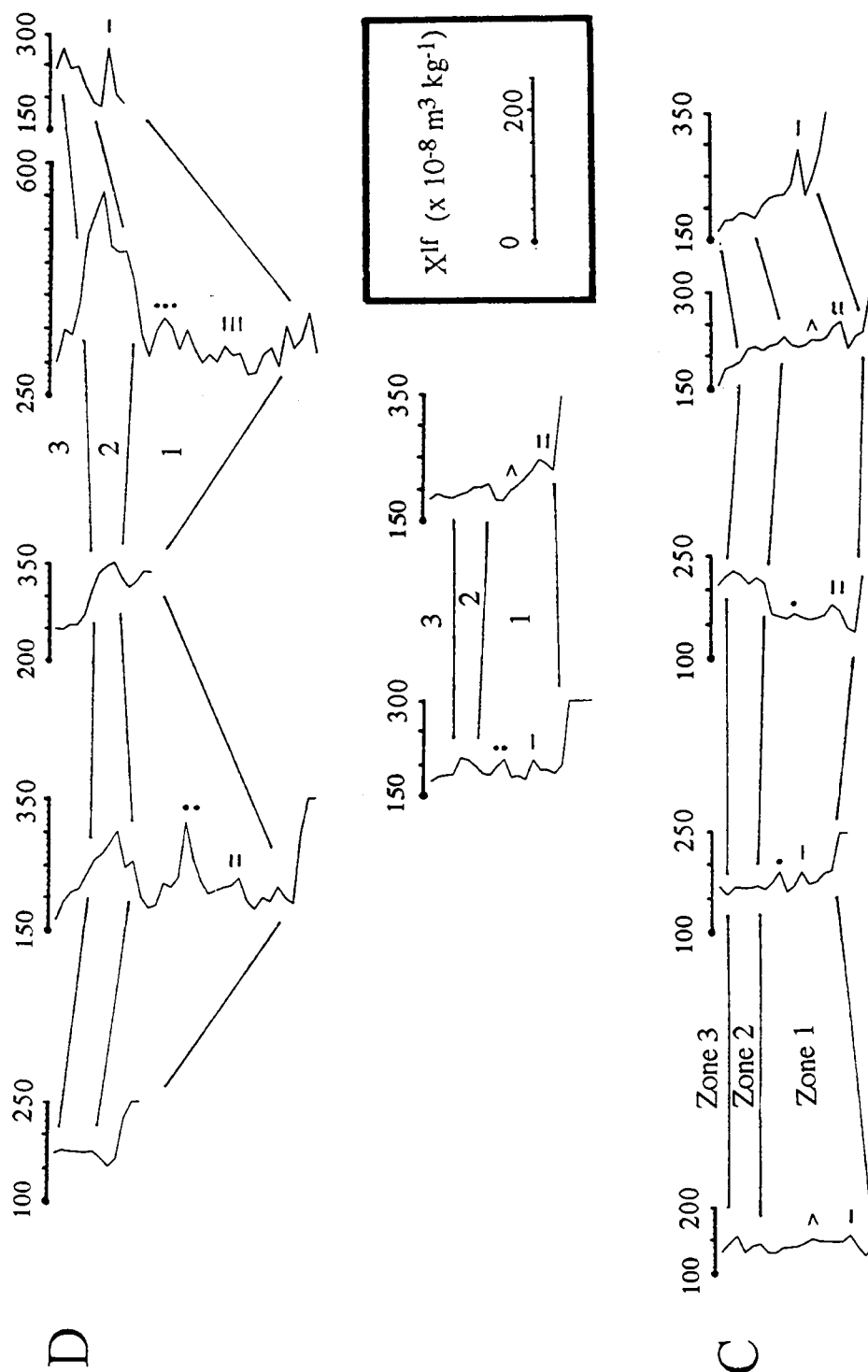
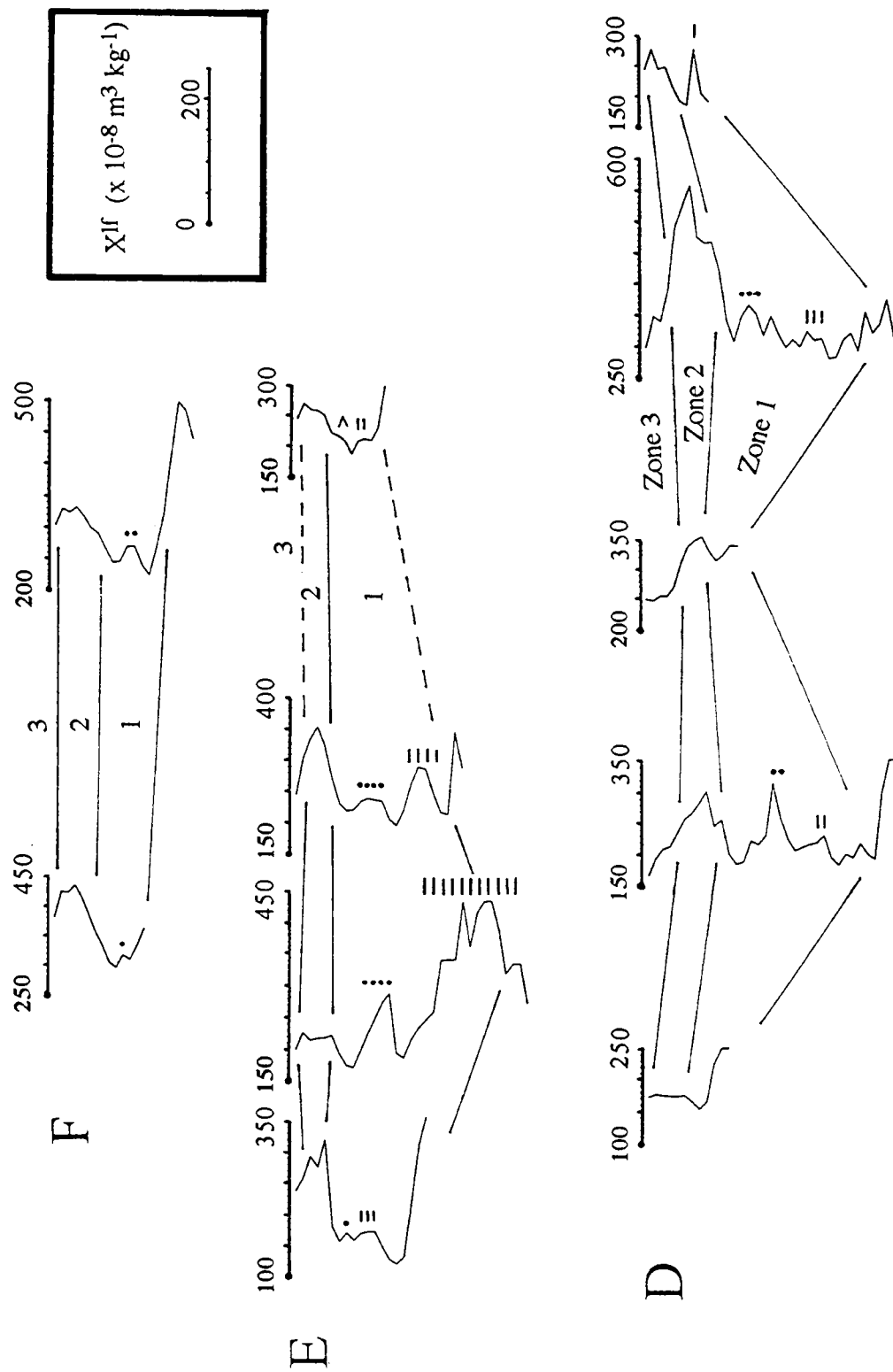


FIGURE 10 (continued)



peaks are shown on Figure 10. Zone 2 features very high susceptibility, especially in core D1 where the peak value is greater than $550 \text{ m}^3 \text{ kg}^{-1}$, the highest value recorded for any sample at the site. The twin-peak nature of deep-water zone 2 is recognizable in deltaic zone 2 as either a lower peak and depression attached to the initial rise of the high peak (e.g., D1 and D3), or as two small but equal-sized peaks at the main peak crest (F3.4, F1.8, E3.8, E4.5). Zone 3 in deltaic profiles is represented by declining susceptibility values to the sediment-water interface, and is thicker in D transect than transects (E or F) closer to the lake inlet. This may reflect scouring conditions at the aggraded lake bed surface in the delta area during recent high flows. This explanation is not, however, favored by the occurrence of low susceptibility values only 6 to 10 cm lower in deltaic cores. An alternative explanation is that sediment entrained by recent flows has been too fine-textured to be deposited near the delta, and instead was washed further out leading to the observed thicker zone 3 accumulations in transect D.

Two cores located halfway between transects C and D, separating deltaic from deep-water areas, were obtained to strengthen correlations between the two environments. Of the two C/D cores, C/D 2.5 provides the strongest link, clearly showing early narrow susceptibility peaks similar to those in C transect zone 1, and a third broader peak (zone 2) forming an overall profile similar to that of cores D3 and E3.2.

Sediment Mass

To calculate mineral mass within each zone, two types of information are needed: 1.) sediment volume in each zonal layer, and 2.) mineral mass per cubic centimeter volume for each layer. Isopach maps showing the thickness of sediment accumulated within each

magnetic zone are presented in Figure 11. Thickest accumulations from zone 1 occur off the distal end of the delta south of the island (Fig. 11a). Zone 2 sediments are more evenly distributed across the lake (Fig. 11b). Zone 3 sediments are generally thickest in the deep-water areas, and thinnest in cores close to the lake inlet (Fig. 11c). Thick zone 3 accumulations occur along the D transect east of the island, whereas little zone 3 sediment occurs south of it. It is possible that this pattern indicates a past shift in position of the main delta distributary. Currently, most flow from the inlet flows north of the island, perhaps explaining this spatial trend in sedimentation.

I calculated the volume of sediment within a magnetic zonal layer by subtracting the volume of overlying water and sediment layers from the volume calculated by including the layer. All volumes were integrated using SURFER software (Golden Software Inc., 1994). The volumes for zones 1, 2, and 3 respectively are 1758, 945, and 532 m³. Unless otherwise contoured, the zero isopach is assumed to be at the shoreline. However, the precise location of the zero isopach is impossible to determine without a much denser coring grid. The total mass error associated with the uncertain location of the zero isopach is minimal because of the small value of sediment thickness for all zones along the lake margin. Variations of a few centimeters in the elevation of the original lake bed surface are observable in adjacent cores at master coring locations, and represent another source of error in sediment thickness, particularly for zone 1. If coring sites randomly sample ridges and furrows in the initial pre-depositional lake bed, 3 to 5 cm in depth, this error should be minimal.

The simple bathymetry of Thompson Lake (Fig. 3b) suggests that thickness of sediment accumulation does not vary greatly over short distances, minimizing zonal interpolation error. If substantial undulation of the lake bed does occur over the 17.5 m distance between depth sounding points along north-south transects, straight line interpolation of zonal thicknesses across these undulations would cause underestimation

FIGURE 11. Isopach maps for sediment zones and event layers. (a.) Zone 1, (b.) Zone 2, (c.) Zone 3, (d.) Event S1, (e.) Event S2.

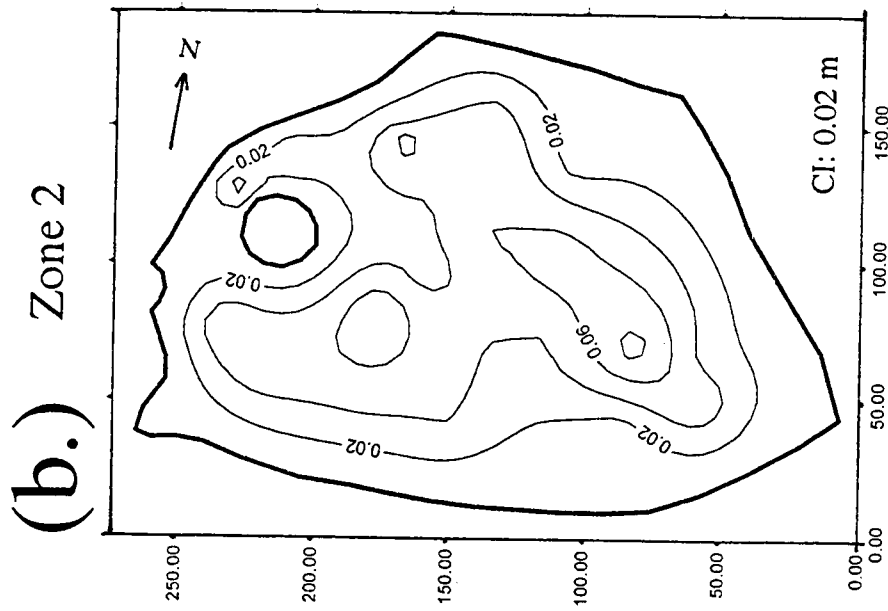
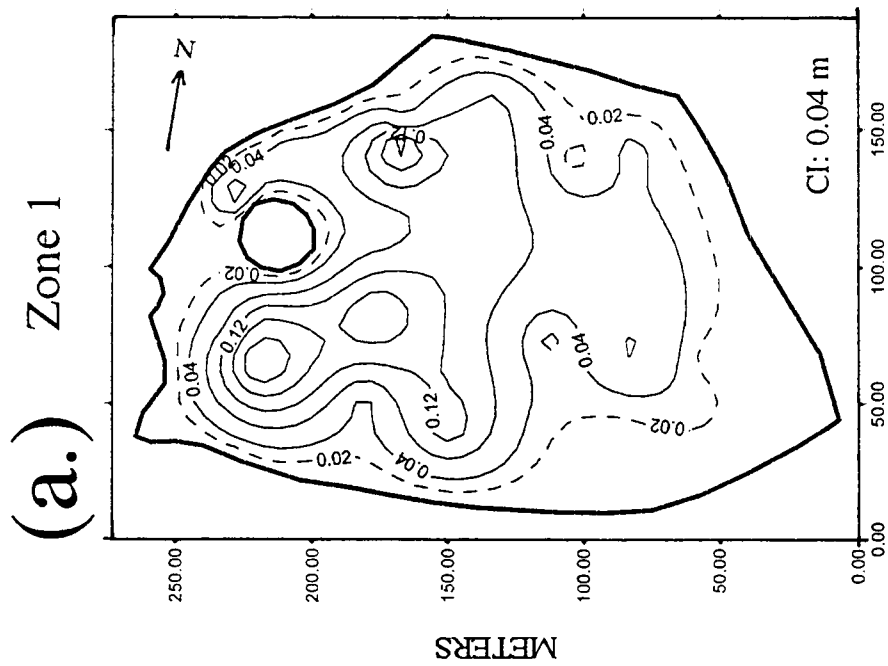


FIGURE 11 (continued)

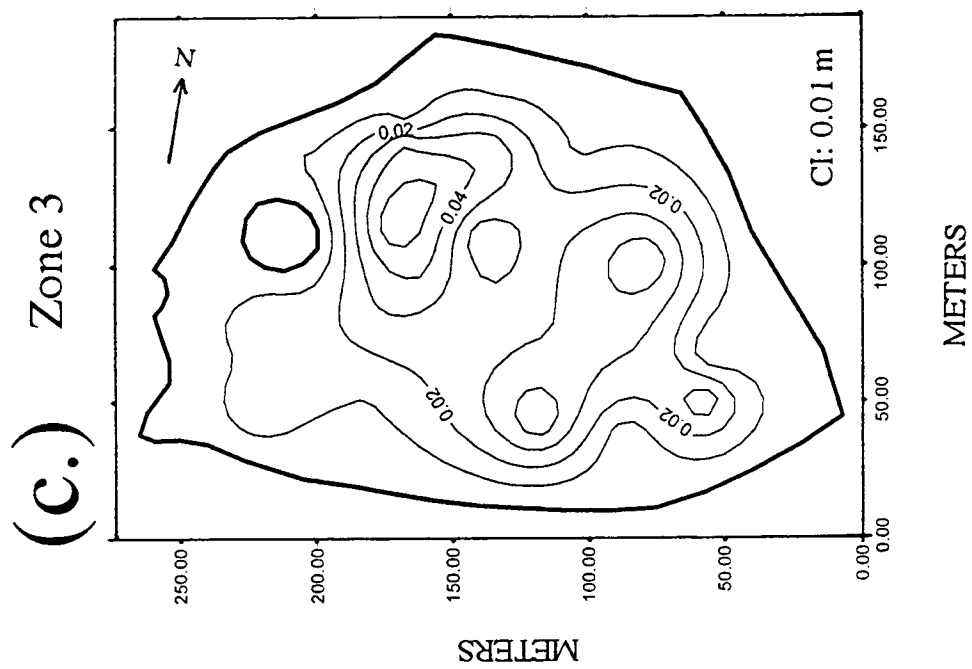
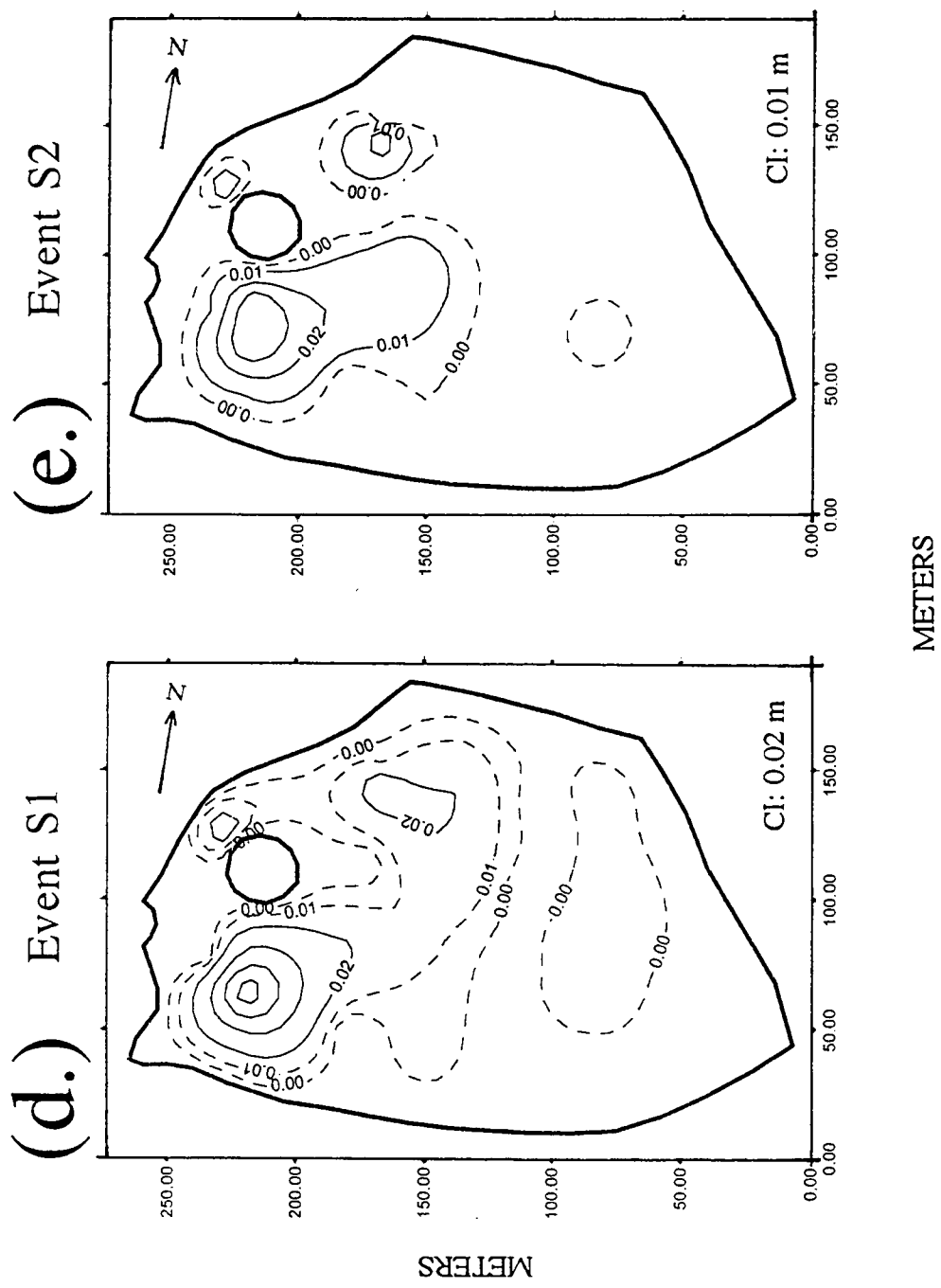


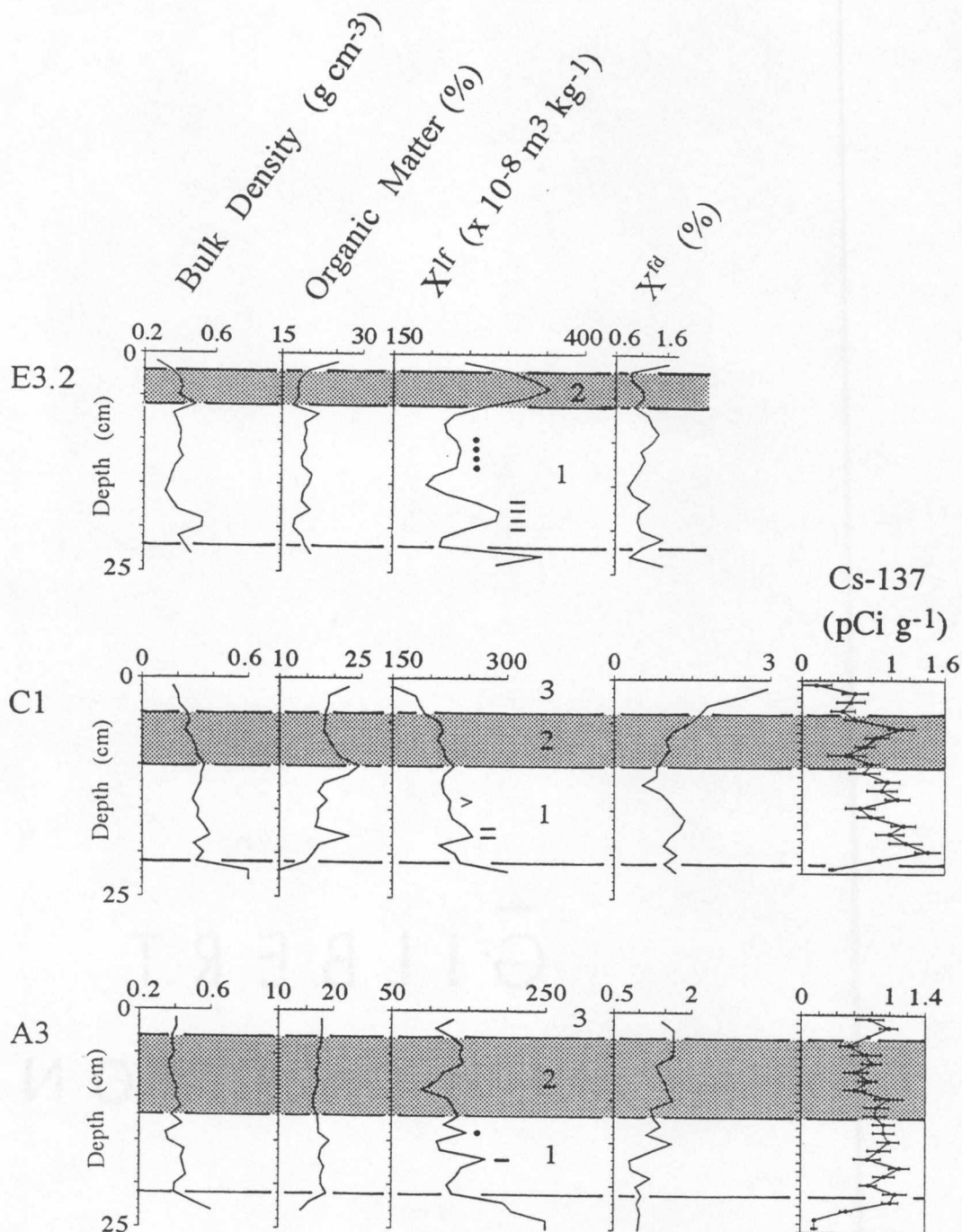
FIGURE 11 (continued)



of zonal volume. The percent error associated with the presence of one, two or three undulations over 17.5 m would be, for a 0.3 m undulation amplitude, less than 1% underestimation in all cases. For a 1-m amplitude, a likely maximum value at Thompson Lake based on valley bottom topography upstream of the reservoir, approximately 1%, 3% and 6% underestimations of volume for one, two and three undulations respectively, would occur. If the common engineering practice of excavating (bulldozing) the reservoir bed to provide materials for the earthen impounding structure was employed, the original lake bed may have been more even than upstream topography, reducing this error to a low value. A rapid decrease of radionuclide levels at the soil interface of sediment cores (described and displayed later; see Figures 12 and 13) suggests prior removal of radionuclide-rich topsoil, perhaps by excavation.

Calculation of mineral particulate sediment mass (bulk density of the mineral sediment) for each magnetic zone is based on loss-on-ignition data obtained for the three master cores (Fig. 12) and assumes negligible contributions of biogenic silica. Despite the differences in depositional environment represented by the three cores, organic percentages derived from loss-on-ignition analyses are similar (range: 16-25%; typical value: 17-19%) in all cores. However, bulk density values are typically higher for deltaic cores because of their higher coarse mineral fraction. Core E3.2 has typical bulk density of 0.4 g cm^{-3} , core C1 is around 0.3 g cm^{-3} , and core A2 (not shown) is around 0.2 g cm^{-3} . Master core A3 (at $\sim 0.4 \text{ g cm}^{-3}$) does not follow this trend because LOI samples were taken after some dewatering had occurred in storage. The loss of water increases bulk density measurements by increasing the contribution of mineral and organic matter (having higher specific gravity than water) to bulk density. Original bulk density values for this core were probably similar to those from core A2, which was analyzed before dewatering could occur. Core A2 LOI data are used in place of data from core A3.

FIGURE 12. Master core data. Zones 1, 2 and 3 are delineated and labelled. Event strata are marked by dashes (event S1) and dots (event S2) as on Figure 10.



Bulk density varies both between zones (i.e., through time), and within a given zone across space in association with compositional effects. Table 4 shows mineral bulk density values derived from the master core profiles, that were used for mass calculation. The greater abundance of large mineral particles accumulating near the delta increases bulk density values there. To account for the intra-zonal variability of bulk density in space caused by changes in mean particle size from delta to deep water, I assigned weighting coefficients to be applied to bulk density values from each master core location using the isopach maps for each zone. Assigned weighting coefficients (percentages) for zone 1 (in order of master cores E3.2 / C1 / A2(A3)) are 0.50, 0.33, 0.17. This somewhat arbitrary breakdown of weighting coefficients reflects the fact that the volume of zone 1 sediment is greatest near master core E3.2 (near the delta), less important near core C1, and still less important near deep-water cores A2 and A3. I assigned weights for zone 2, having relatively even distribution of sediment across the lake, of 0.33, 0.33, 0.33; for zone 3 with maximum accumulation near core C1, the values are 0.17, 0.50, 0.33. With these numbers, a representative value of bulk density for each zone can be calculated by multiplying each zonal bulk density from the three master cores by its weighting coefficient reflecting the location of the core. Resulting values are: zone 1 = 0.28 g cm^{-3} ; zone 2 = 0.24 g cm^{-3} ; zone 3 = 0.19 g cm^{-3} . Multiplying these values times zonal volumes results in zonal mineral masses (metric tons, t) of : zone 1 = 492.2 t ; zone 2 = 226.8 t ; zone 3 = 101.1 t. Total mineral mass for all accumulations = 820.1 t. The computation of sediment yield from this mineral mass is made by subtracting the lake bank input (2.9 t yr^{-1} times 29 years) and airborne particulate input (0.0332 t yr^{-1} times 29 years), each multiplied by 0.81 to correct for trap efficiency, correcting the result for trap efficiency by dividing by 0.81, then dividing the result by 29 years and 3.83 km^2 catchment area. The result is an average 29-year sediment yield of $8.3 \text{ t km}^{-2} \text{ yr}^{-1}$.

TABLE 4. Mineral Bulk Density Values (g cm^{-3}) Used in the Calculation of Zonal Sediment Mass.

Master Core	zone 1	zone 2	zone 3	event S1	event S2
E3.2	0.32	0.34	0.27	0.38	0.32
C1	0.27	0.24	0.18	0.28	0.23
A2(A3)	0.18	0.17	0.16	0.22	0.22
weighting (E3.2/C1/A2(A3))	.5/.33/.17	.33/.33/.33	.17/.5/.33	.5/.33/.17	.5/.33/.17
weighted bulk density	0.28	0.24	0.19	0.32	0.27

Storm Events and Sediment Yield

Many core profiles display high-magnitude but short-lived variations in X^{lf} that occur over an otherwise continuous trend in susceptibility. Two such variations (events S1 and S2) have been identified within zone 1 and traced from core to core across the lake. Of the two events, the first is represented by higher X^{lf} peaks and is more clearly visible in most of the cores (Fig. 10). I interpret these X^{lf} peaks to represent deposition from individual large storms. The evidence supporting this interpretation is: 1.) X^{lf} fluctuations used to identify the events represent particle-size increases; 2.) peaks rise, and also often fall, abruptly, and occur over an interval otherwise characterized by low X^{lf} values; 3.) isopach maps for these layers show thick coarse accumulations in the deltaic area; small events transport finer sediment more easily held in suspension that would be less likely to settle to the lake bed close to the lake inlet; 4.) if these event layers represent moderate floods or even low flow periods, this would mean that the Agnes event, a very widespread climatic event representing the largest precipitation event by far at the site, produced no observable effect on the sediment record. It is unlikely that a moderate event produced a large effect on X^{lf} whereas a known catastrophic event produced no observable effect.

Precipitation data from Mt. Weather, and discharge data from Goose Creek (Fig. 5), indicate moderate to low maximum flows from 1965 to 1972, the year of the Hurricane Agnes storm. The Agnes event would have produced heavy flooding in the Thompson Lake vicinity because of high precipitation totals and resulting high discharge as observed at Goose Creek and most other stations in central and northern Virginia. High discharge associated with Hurricane Agnes may have been the cause of coarse particle input reflected in the X^{lf} record as the initial isolated peak of event S1. The

second peak (event S2) may similarly be related to a second large precipitation and discharge event occurring in 1977 (Fig. 5).

Assuming that these X^{lf} events do in fact occur during years in which catastrophic flooding occurred, it is difficult to separate out sediment contributions from the large event from the presumably smaller but important fraction associated with moderate and small events occurring during the same year. Within a one-centimeter slice of sediment (the sampling interval of this study), the coarse sediment contribution of a large event may only represent a fraction of the entire one-centimeter thickness and still increase X^{lf} values. In the correlation diagram (Fig. 10), I assigned event layer thicknesses by assuming that high, sharp one-centimeter-thick peaks represent flood dominance of sedimentation (i.e., full-peak thickness is flood sediment) and subdued one-centimeter-thick peaks represent a smaller contribution of flood sediment (i.e., 0.5 cm thickness is flood sediment) that is diluted by the year's total sediment yield. Figure 11d shows isopach maps resulting from the correlation and layer thicknesses displayed in Figure 10. Event S1 is associated with thickest accumulations in the delta region south of the island. Total volume for event S1 accumulation equals 255 m^3 . Mineral bulk density in the event S1 layer is higher than the average for zone 1, in all three master cores. The estimated values for each master core are: E3.2 = 0.38 g cm^{-3} , C1 = 0.28 g cm^{-3} , and A2 (A3) = 0.22 g cm^{-3} . Because this sediment is concentrated in the delta region, and increasingly less so towards deep water, I use the weighting coefficients 0.50, 0.33 and 0.17 to calculate bulk density values of event S1 using the three master cores E3.2, C1 and A2 (A3) respectively. Average mineral bulk density using this weighting equals 0.32 g cm^{-3} . For a volume of 255 m^3 , this value of bulk density results in a total mass of 81.5 t, about 10% of the total 29-year accumulated mineral mass of 820.1 t. Similar calculations are accomplished for event S2 (volume = 137 m^3) using the values for cores E3.2 / C1 / A2 (A3) respectively: mineral bulk density = $0.32 / 0.23 / 0.22 \text{ g cm}^{-3}$; weighting

coefficients: 0.50, 0.33, 0.17. The resulting value for average mineral bulk density is 0.27 g cm^{-3} , and mineral mass for event S2 is 37 t, or 4.4% of the 29-year accumulation.

Chronology: Cs-137

Calculations of sediment yield derived from sediment masses in zones 1, 2, and 3 require dates to be placed at stratigraphic contacts. The combination of Cs-137 and Pb-210 stratigraphy provides the chronology for sediment at Thompson Lake. Cesium-137 (half-life = 30.2 yr) accumulating in lake sediment is ultimately derived from airborne fallout produced during periods of nuclear testing. Levels of fallout during these periods are well documented for the northern hemisphere, and described and published in federal research lab documents as well as a number of professional papers (e.g., Ritchie et al., 1973; Ritchie and McHenry, 1990; Foster et al., 1990; Walling and He, 1993). These records all record the same basic trends: the initial fallout Cs-137 rise in 1954, an initial peak in 1958, decline and subsequent rise again to a maximum peak in 1963, and gradual decline from 1963 to near zero levels in 1973. Fallout associated with the Chernobyl disaster in 1982 was marked in Europe (Rowan et al., 1993) and occurred in low but measurable amounts in the US (Lauren Larsen, personal communication). Sedimentation at Thompson Lake began at latest in summer 1966 during the period of gradual decline of fallout Cs-137 to low levels in 1973. Because the low levels of Chernobyl fallout are not likely to be recognizable in lake sediment from the US, the most useful Cesium time marker at Thompson Lake is the attainment of low values in 1973.

Fallout Cs-137 accumulates in topsoil through adsorption onto fine soil particles. Erosion and redeposition of this topsoil constitutes a second input of Cs-137 into lake

sediment. The Cs-137 profile through a sediment core reflects the fallout component modified by the loss through outflow, and dilutions or additions provided by topsoil redeposition (Ritchie and McHenry, 1990). Loss through outflow is assumed to be roughly constant. The effect of topsoil redeposition can be estimated by looking at the near surface Cs-137 profile and total Cs-137 inventory of lake sediment. High topsoil erosion rates during high fallout periods tend to maintain the fallout pattern in downstream lake sediment accumulations, particularly if sediment delivery ratios are high.

Average Cesium levels in core A3 (Fig. 12) are highest in basal sediments of magnetic zone 1, and generally decline up-core. The Cesium decline to upper sediments is not monotonic however. Zone 1 is characterized by two peaks separated by lower values, followed by a rise to moderate levels continuing to zone 2. Zone 2 begins with a slight drop in activity preceding a peak of activity similar to that of late zone 1, followed by a sharp decline to relatively low values. Cesium-137 activity rises at the beginning of zone 3. Values never reach zero in the upper core but instead fluctuate about a low average associated with topsoil- or Cs-137-rich channel sediment input. This general pattern is replicated in the coarser sediments of core C1 (Fig. 12), with a few differences. Initial high values at the base of zone 1 decline steeply to low activity midway through the zone. A subsequent sharp rise is followed by slow consistent decline to the beginning of zone 2. This gradual decline of moderate values matches the interval of stable moderate values in core A3. The small rise at the end of this interval (in core A3) early in zone 2 is matched in core C1 by a large rise and subsequent rapid decline of Cs-137. The importance of erosional Cs-137 can be seen by comparing Cs-137 inventories for lake sediment (2.63 pCi cm^{-2}) to that of fallout since 1965 (1.8 pCi cm^{-2} ; measured at New York City). The difference (0.83 pCi cm^{-2}) represents erosional Cs-137 contribution. Very low Cs-137 values in soil underlying lake sediment indicate either high levels of

pre-1965 topsoil erosion, or more likely, disturbance of topsoil during (and possibly for the purpose of) reservoir construction. Variability associated with counting error (note the width of error bars on activity plots), and erosional inputs obscure the post-1963 monotonic decline in fallout to near-zero levels.

Erosional Cs-137 inputs depend on the source of sediment (Cs-rich topsoil vs. Cs-poor subsoil vs. (Cs-undefined) in-channel) and the particle-size attributes of transported material (Cs-137 adsorbs more readily on finer particles). Because streambank erosion appears to be an important source for newly eroded soil (see monitoring chapter), Cs-enriched topsoil may be a relatively unimportant source of Cs-137 for lake sediment. Sediment within the channel during the 1950s and 1960s would have become enriched from fallout, especially given the dispersed nature of surface flow (and hence large channel area) through boulder fields in upper regions of the catchment. Accordingly, streambanks and channel sediment may be the low- and high-Cesium sources mixed to produce the lake sediment signal after the high fallout period. If so, the travel time of sediment in active storage to the lake, must be greater than 30 years since Cs-137 input continues today. Alternatively, Cs-137-enriched topsoil could be the source of mineral sediment measured at the knickpoint (noted in chapter 5) and therefore might provide some of the Cs-137 currently accumulating in Thompson Lake. The strong affinity of Cs-137 with fine-grained sediment means that small to moderate flow events preferentially winnowing small particles from stored channel sediments could produce higher Cesium values (per unit mass) in lake sediment. Alternatively, large floods may be associated with increased topsoil erosion from sideslopes and the debris train surface where infrequent erosive overland flow allows only gradual depletion of adsorbed Cesium. These events would entrain both fine and coarse particles, and later, with differential settling rates, produce graded layers of sediment with uppermost portions enriched.

Coherent fluctuation in magnetic susceptibility indicative of storm layers in cores A3 and C1 is not matched by coincident positive or negative fluctuation in Cs-137 levels. At Thompson Lake, there is no clear and consistent relationship between magnetic susceptibility (a surrogate of particle size at Thompson Lake) and Cesium values for cores with Cesium data (A3, C1; Fig. 12). Without a clear relationship between storms and Cs-137 content, or a Cs-137 profile closely matching the fallout timeline, Cs-137 profiles are probably most useful as indicators of relative rather than absolute sedimentation rates. I interpret high but coherent (rises and declines cover more than one centimeter) fluctuation in early zone 1 to represent overprinting of the fallout signature by flood events removing the most enriched sediments and near channel soils that had accumulated high activities during the immediately prior period of high fallout in 1958 and 1963/64. The plateau of late zone 1 and the slight decline in early zone 2 represent relative stability in sediment influx possibly produced by a succession of moderate events over several years. Low values in upper zone 2 reflect continuing depletion of Cs-137 within the channel due to removal of older enriched sediment and radioactive decay without replenishment, and/or generally low sediment input during this time of almost zero fallout. The presence of a moderate increase in Cs-137 during zone 3 of both cores A3 and C1 suggests increased erosional input, since fallout remains low.

This interpretation of the Cs-137 records is consistent with precipitation and discharge trends observed in data from Mt. Weather and Goose Creek (Fig. 6). The record of annual precipitation shows initially low precipitation in the late 1960s, increases in the 1970s with a high peak from Hurricane Agnes in 1972, followed by a series of moderately high values in the late 1970s that may account for the Cs-137 "plateau" of middle to late zone 1. This wet period is followed by the 1980s dry period, possibly accounting for generally low Cs-137 values in late zone 2 as a result of low erosion rates. This dry period is broken up by one large flood in 1984, possibly coinciding with the

peak in Cs-137 at the end of the Cs-137 plateau. Wetter conditions in the early 1990s may correspond to recent Cs-137 increases. If this correlation is correct, the initial rise of the Cs-137 plateau would occur around 1975 when high annual precipitation and discharge maxima become more common. The Cs-137 peak at the end of the plateau in zone 2 would occur at 1984, and the zone 3 Cs-137 rise would represent the year 1992.

Chronology: Pb-210

The accumulation and transport of Pb-210 (half-life = 20.3 yr) in the environment is different from Cs-137 in two important ways. First, Pb-210 is a natural radionuclide that is a continuous component of atmospheric fallout; its fallout rate can be assumed to be relatively constant for a given area. Second, Pb-210 can also be generated *in situ* in lake sediment through the decay of Radium-226 (half-life = 1600 yr) in the sediment. Lead-210 chronology is based on the separation of the atmospheric fallout component (termed "unsupported" Pb-210) from the *in situ* component ("supported" Pb-210), and the decay rate of the supported component. Two main categories of models describing the occurrence of Pb-210 in lake sediments exist. Constant initial concentration (CIC) models apply when the accumulation rate of lake sediment is nearly constant. In this case, Pb-210 will decrease exponentially down-core in accordance with its decay constant ($\ln 2 / \text{half-life}$). Constant initial concentration models are inappropriate at Thompson Lake, where fluvial inputs clearly provide changing conditions of sedimentation through time. Under these conditions, constant rate of supply (CRS) models which assume constant rate of fallout variably diluted by changing input of Pb-210-poor sediment are preferred (Appleby and Oldfield, 1978). Both models require the input of unsupported Pb-210 values so that this component must be separated from the total Pb-210. The supported

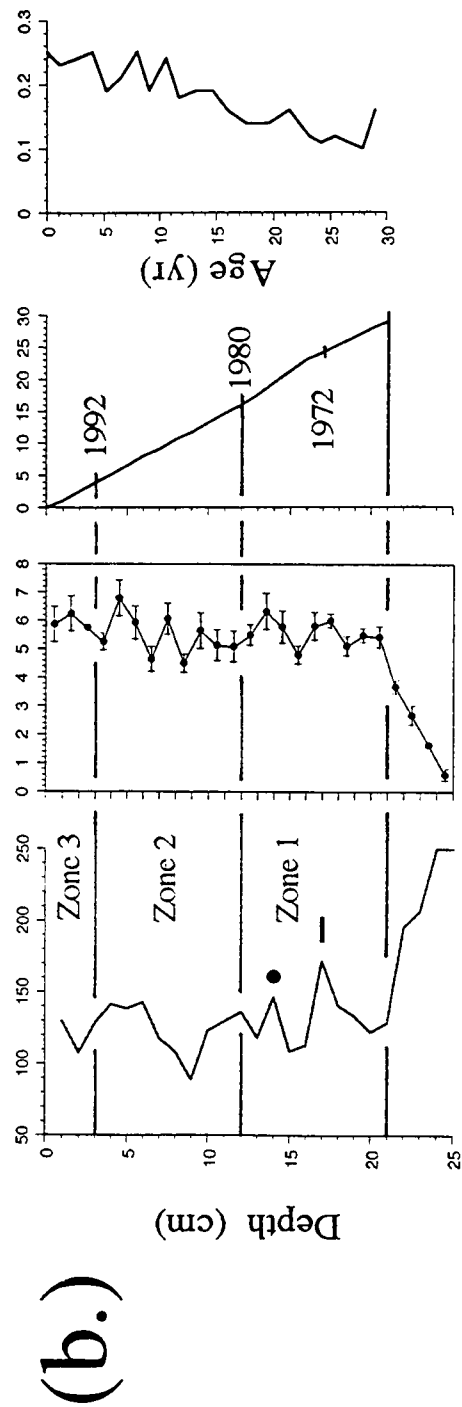
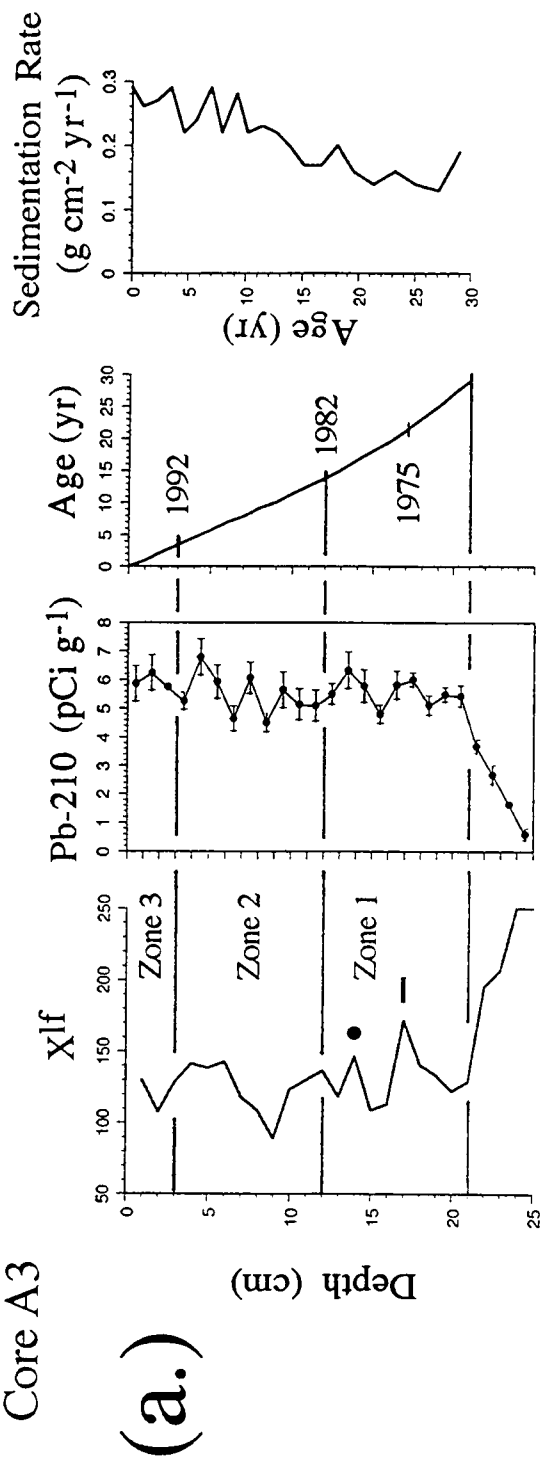
Pb-210 can be quantified by measuring the abundance of Pb-214 (half-life = 26.8 min.) derived from decay of sedimentary Radium-226, and in equilibrium with its decay product, supported Pb-210 (Cutshall et al., 1986). At Thompson Lake, the supported component was found to be too low to measure, so that all Pb-210 activity measurements in Thompson Lake sediments represent the unsupported component.

The Pb-210 profile for core A3 is shown in Figure 13 (a and b). The profile is less erratic than Thompson Lake Cs-137 profiles, in part because counting error is lower. Also, as mentioned earlier, Pb-210 fallout is roughly constant regionally, stabilizing the overall signal. The profile does not display monotonic decline according to the decay constant of Pb-210, confirming that the CIC model is inappropriate at this site. Fluctuation in Pb-210 values indicate that the fallout signal is altered by sediment inflow and therefore the CRS model is deemed more appropriate. The CRS model assumes that the base of a sequence is old enough that little unsupported Pb-210 remains. Dates are then based upon cumulative values of unsupported Pb-210 in relation to the total inventory (Appleby and Oldfield, 1978). The 29-year basal age of Thompson Lake sediment is insufficient for decay to have removed this unsupported Pb-210. I adjusted the model by treating the 29 years of sediment as if it were the top of an older profile and calculating the total unsupported Pb-210 that would be required to satisfy the 29 year age of the sediment. I added this total to values of unsupported Pb-210 obtained for Thompson Lake sediments and calculated sedimentation rate and age using the CRS equations (see Appendix). The application of the CRS model to this profile gives a curve of age vs. depth, and a curve showing the associated changes in sedimentation rate through time (Fig. 13).

Two chronologies are presented in Figure 13, the first (Fig. 13a) based solely on Pb-210 values, and the second (Fig. 13b) based on Pb-210 plus an assumed date of event stratum S1 (Hurricane Agnes?) of 1972. The first age-depth curve (Fig. 13a) places the

FIGURE 13. Pb-210 profile and chronologies from master core A3. (a.) Chronology derived solely from the Pb-210 profile; (b.) Chronology in (a.) adjusted for the probable 1972 (Hurricane Agnes) date of event S1 Event strata are indicated by dashes (event S1) and dots (event S2) as used in Figure 10.

Core A3



date of the zone 1 / zone 2 boundary at mid-1982; the zone 2 / zone 3 boundary is dated at early 1993. Greatest variability in sedimentation rate occurs in the upper 10 cm of core A3. Peaks in sedimentation rate occur in 1992, early 1989 and early 1987, with smaller peaks in the lower core at 1977 and 1972. The two peaks in magnetic susceptibility notable in magnetic zone 1 (events S1 and S2) are dated by this model at early 1975 and early 1980. The second age-depth curve (Fig. 13b) places the zone 1 / zone 2 boundary in 1980 and places the zone 2 / zone 3 boundary in 1992. Greatest variability in sedimentation rate occurs, as in Figure 13a, in the upper 10 cm of core A3, with peaks in 1991, 1987, and early 1985, with smaller peaks in early 1974 and early 1970. Events S1 and S2 occur respectively in the years 1972 (by definition) and 1977. Both Pb-210-based age-depth curves are consistent with the interpretation of Cs-137 profiles discussed earlier.

The CRS model assumes that fallout Pb-210 is diluted by input of sediment not enriched by adsorption of this fallout. Observation of processes within the catchment suggest that topsoil input, the main source of enriched sediment, is of minor importance compared to subsoil input through bank collapse. This observation is not surprising considering that topsoil erosion is usually less important in fully forested areas and more so in agricultural or deforested areas (Appleby and Oldfield, 1983). The average sedimentation rate for the 29-year period and the trap efficiency of Thompson Lake fall within acceptable limits as theorized by Dearing and Foster (1993) for use of Pb-210. Marked fluctuation in magnetic susceptibility and radionuclides within 1-cm intervals and within relatively continuous "background" trends suggests that lake sediment bioturbation is relatively ineffective below 1-cm at this site. These site characteristics indicate that, of the dating methods available, Pb-210 offers the best means of providing accurate chronology for core A3, and sedimentation across the lake.

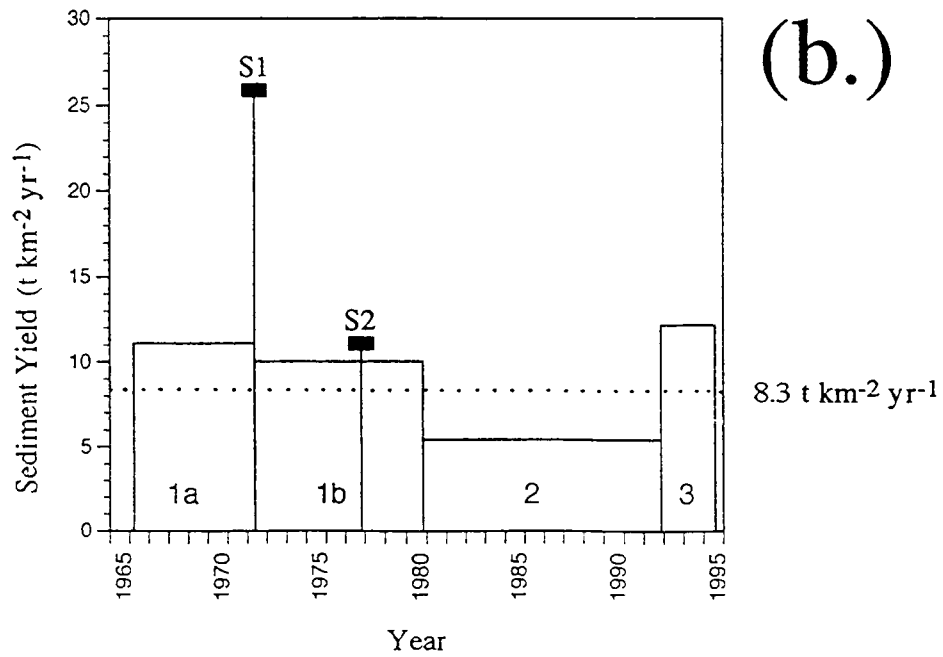
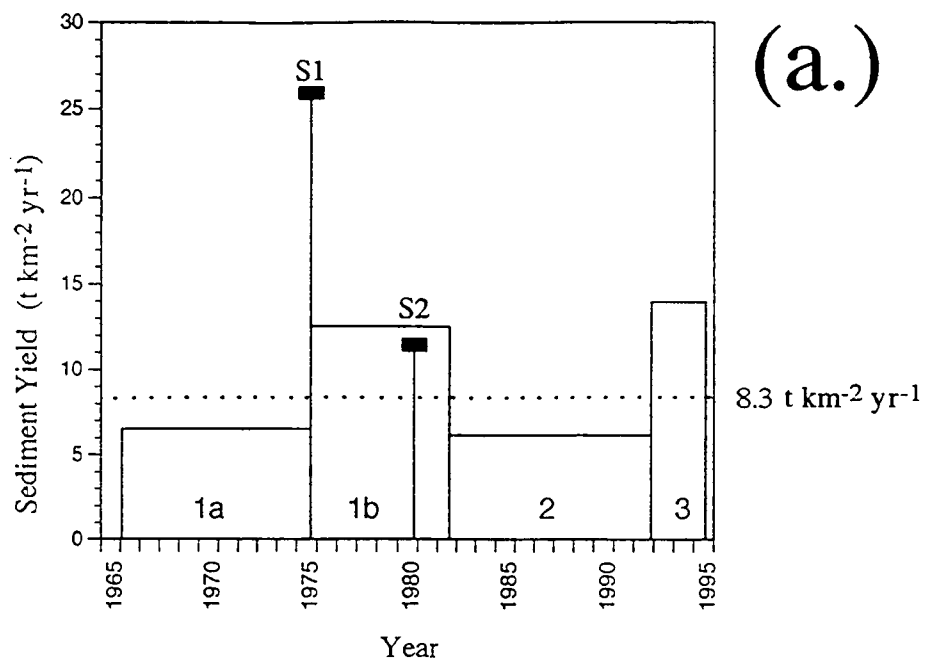
Sediment Yield

Catchment sources of sediment accumulating in Thompson Lake are erosion in unspecified areas along first-order streams during floods and erosion of subsoil exposed in channel banks primarily along the second-order channel. These sediments have magnetic susceptibilities related to their particle-size attributes and therefore to flow competence. Three major changes and two minor fluctuations (events S1 and S2) in susceptibility are traceable between cores and define magnetic zones for which sediment volume and mass have been calculated. Lead-210 appears to provide the best chronology for these zones. Dividing sediment mass in each zone by the time of deposition indicated by ages of zonal boundaries and the catchment area gives sediment yield.

Sediment yield for each zone is plotted in Figure 14. This figure shows calculated sediment yield for time periods defined by Pb-210 dating of zonal boundaries and events S1 and S2 horizons. Two scenarios are presented (Figs. 14a and 14b) that are associated with the two different chronologies of Figure 13. The first (Fig. 14a) represents the Pb-210 chronology produced using only the basal date (1966) as an independent time marker. The second (Fig. 14b) assumes that the event S1 stratum is associated with Hurricane Agnes and therefore is an independent time marker of the year 1972. Average 29-year sediment yield of $8.3 \text{ t km}^{-2} \text{ yr}^{-1}$ is shown by a horizontal dotted line.

In the first scenario (Fig. 14a), the basal stratum, zone 1, is divided into two subzones (1a and 1b) of volumes 642 and 861 m^3 respectively by event S1 sedimentation. I did not further subdivide subzone 1b using event S2 because this event is not as well expressed as event S2 in all parts of the lake, reducing my confidence in its correlation. Mineral masses for subzones 1a and 1b calculated using the weighted bulk density for zone 1 (Table 4), correcting for lake bank erosion and airborne particulate input, and adding one-half of the event S1 mass (81.5 t) to each, are 198.0 t and 265.5 t

FIGURE 14. Sediment yield diagrams. Average 29-year sediment yield ($8.3 \text{ t km}^{-2} \text{ yr}^{-1}$) is shown by the horizontal dotted line. (a.) Chronology based solely on the Pb-210 profile; (b.) Chronology based on the Pb-210 profile assuming that the event S1 peak occurs in 1972 (Hurricane Agnes).



respectively. (Note: mineral masses for these subzones must incorporate the corrections for lake bank erosion and airborne particulates (unlike the raw mass values given below for the other zones), because they include the event S1 mass component which is unaffected by these inputs, due to its short time interval of deposition.) Sediment yield is calculated for each zone as described earlier for the total 29-year accumulation, except that the deposition time for each zone is substituted in place of 29 years. Average sediment yield was low ($6.7 \text{ t km}^{-2} \text{ yr}^{-1}$) during the latter half of the 1960s (zone 1a), high ($12.5 \text{ t km}^{-2} \text{ yr}^{-1}$) through the 1970s (zone 1b), low ($6.2 \text{ t km}^{-2} \text{ yr}^{-1}$) during the 1980s (zone 2), and high ($13.8 \text{ t km}^{-2} \text{ yr}^{-1}$) in the early 1990s (zone 3). Applying the 81% trap efficiency correction to event S1 mass (81.5 t) and assuming that this event in fact does primarily represent sedimentation from a rapid depositional event related to a large storm, the resulting event sediment yield estimate of 26.3 t km^{-2} represents 10.9% of the total 29-year catchment sediment yield, and is 3.2 times the annual average. The rationale for events S1 and S2 to represent sediment layers from isolated storm events has been given earlier.

In scenario 2 (Fig. 14b), the chronology is recalculated holding the date of event stratum S1 at 1972, assuming that this stratum represents the Hurricane Agnes event. This assumption is supported by the radionuclide chronology, as well as by the relative timing of storm events in the precipitation and discharge records. The Cs-137 profiles place zone 1 sedimentation in the 1970s and the Pb-210 depth-age curve places the initial X^{lf} rise of event S1 between 1972 and 1975. Also, event S1 is the first positive fluctuation in X^{lf} in the record, and Agnes was the first major storm in this area in the precipitation and discharge records following the dry period of the late 1960s (Fig. 5). The second event (S2) falls at 1980 on the depth-age curve from Pb-210. The second largest one-day precipitation event at Mt. Weather over the last 30 years occurred on August 24, 1979 (14.1 cm), and was only slightly smaller than the Agnes event (at 19.4

cm). Discharge downstream, however, is only moderately high at this time (Fig. 5), indicating either a lower discharge response due to, for example, dry antecedent conditions, or, that this precipitation level was inconsistent across the Goose Creek watershed. A moderately large precipitation event that is also associated with high downstream discharge occurred in 1977 and could be an alternative cause of event S2, assuming that the Pb-210 chronology is in error by three years. Recalculation of the Pb-210 chronology based on a 1972 age for event S1 places the occurrence of event S2 during 1977. Using this adjusted chronology, calculated sediment yield was high ($11.3 \text{ t km}^{-2} \text{ y}^{-1}$) during the latter half of the 1960s (zone 1a), and remained high ($10.0 \text{ t km}^{-2} \text{ y}^{-1}$) through the 1970s (zone 1b). Sediment yield was low ($5.3 \text{ t km}^{-2} \text{ y}^{-1}$) during the 1980s (zone 2), before reaching its highest value ($12.1 \text{ t km}^{-2} \text{ y}^{-1}$) in the early 1990s (zone 3) (Fig. 14b).

Comparison of these lake sediment-based estimates of sediment yield to estimates derived from extrapolation over time of the few measurements of suspended sediment during baseflow and flood conditions at Thompson Lake reveals similarity. Average baseflow sediment concentration (Table 2) is around 0.004 g l^{-1} (4 t m^{-3}), and average flood sediment concentration for the small to moderate events sampled is around 0.0358 g l^{-1} (35.8 t m^{-3}). Visual observation of stream turbidity at other times, and suspended sediment data (Table 2) show that sediment concentration is more related to the occurrence of precipitation than to discharge alone. Winter baseflow is much higher than that in summer, yet both baseflows have similarly low sediment concentrations. If more than a trace of precipitation occurs on 110 days (30%) of the year, each time producing the average concentration of sediment noted above, the annual value for sediment yield equates to $6.7 \text{ t km}^{-2} \text{ yr}^{-1}$, a value close to the $8.3 \text{ t km}^{-2} \text{ yr}^{-1}$ derived from lake sediment. This calculation assumes (unrealistically) that the estimated $1,900,000 \text{ m}^3$ annual discharge is distributed evenly throughout the year, the average sediment

concentration is maintained (or limited to) the entire day of the precipitation and that bed load is not important during moderate flows. Adding in bed load would increase this sediment yield estimate.

Given the dearth of monitoring data and the resulting assumptions made to enable a calculation, the value of sediment yield derived from limited direct monitoring is useful mainly as an indicator of relative magnitude of the difference produced using the two data types. In this case, even given the unrealistic assumptions for process extrapolation, the values for sediment yield obtained appear to be compatible.

VIII. SYNTHESIS AND DISCUSSION

Overview

The preceding chapter documented the progression of steps taken to produce a record of medium-term sediment yield at Thompson Lake. The two prior chapters dealt with analyses of environmental characteristics and history that serve to provide a check on the reconstruction of sediment yield, and enable an interpretation of it. The purpose of the current chapter is to assess the quality of the sediment yield diagram, to place it within the context of other sediment yield information, and to interpret it on the basis of the environmental characteristics of the basin and its history. In addition, I analyze the error in the reconstruction and its consequences for interpretation.

Medium-term Sediment Yield Estimates

The average annual sediment yield based on 29 years of sediment accumulation at Thompson Lake ($8.3 \text{ t km}^{-2} \text{ yr}^{-1}$) is consistent with predictions by Wolman (1967) of less than $40 \text{ t km}^{-2} \text{ yr}^{-1}$ for the northern Piedmont Province under forested conditions. It is also consistent with the regression relationship between percent forest cover and sediment yield for Potomac drainage sub-basins ($\sim 9 \text{ t km}^{-2} \text{ yr}^{-1}$) for 100% forest cover (Wark and Keller, 1963; reproduced in Gregory and Walling, 1973). Wolman's estimate is derived in part from two catchments in Maryland (areas of 19.2 and 18.9 km^2) with corresponding sediment yields of 4.2 and $1.9 \text{ t km}^{-2} \text{ yr}^{-1}$. These catchments represent the northern extension of the Thompson Lake geology and topography and therefore have

many similarities to the study site. Higher sediment yields are expected from smaller watersheds due to reduced chances for sediment storage and consequent higher sediment delivery ratio (SDR) (Walling, 1983). This spatial characteristic may in part explain low sediment yield values for the larger Maryland streams relative to those at Thompson Lake.

A regression relationship between catchment area and SDR for the southeastern US Piedmont (Walling, 1983) places the SDR for 20-km² forested catchments at roughly 20% less than that for basins the size of Thompson Lake. This adjustment, if applied to the similarly sized Maryland catchments, results in sediment yield values near the low values reconstructed for the late 1960s and 1980s at Thompson Lake. Of course, regression relationships may be based on points that have high scatter, and hence this particular empirical curve may not apply exactly to SDRs at these sites. The 18.9 km² basin (Fishing Creek) has sediment yield much lower than the Thompson Lake average, despite its similar relief and lithology. The Fisher Creek drainage net runs parallel to regional structure, unlike the Thompson Lake drainage net, and perhaps differs in its hydrological response to storms. Alternatively, differences in general precipitation characteristics, precipitation history (data given by Wolman are prior to the period of deposition at Thompson Lake), the length of available record, or sediment control strategy (Fisher Creek is currently, and may also in the past have been, managed for water supply) may be responsible for the noted differences in yield. Order of magnitude similarity between lake sediment-based average sediment yield estimates from Thompson Lake and sediment yield measurements derived from short-term monitoring at Wolman's Maryland sites indicates the general accuracy of the lake sediment-based estimates.

As might be expected, comparisons with basins farther away reveal less similarity. To the north in central Pennsylvania, small (2 km²) catchments heading on the

linear flanks of Blue Mountain have much larger sediment yields (30 to 50 t km⁻² yr⁻¹) (Reed, 1980). These values are probably too high to attribute to higher SDR, according to another regression of basin size vs. SDR presented in Walling (1983) for the eastern US. These basins however, are deforested in lower reaches and are also underlain by relatively weak sedimentary lithologies, perhaps explaining their relatively high sediment yields. In fully-forested and similarly sized Coweeta basins in North Carolina, which are underlain by more resistant gneisses and schists, measured sediment yields are very low (3 t km⁻² yr⁻¹). Velbel (1985) has inferred from these low sediment yields that most long-term sediment yield must involve events of low frequency and high magnitude to account for long-term denudation of the Appalachian Mountains. These Coweeta catchments which serve as control basins for forestry experiments, receive high annual rainfall, and are steep, but may have less disturbed soils than at Thompson Lake, given the relative lack of past habitation and forestry. In addition, they lack the thick, erodible colluvial fill typical of lower reaches of the Thompson Lake watershed.

Sediment yield values from forested Hubbard Brook sub-basins are similar to those from Coweeta. The Hubbard Brook basins are less than 1 km² in area and have sediment yields of only around 2.2 t km⁻² yr⁻¹ (Bormann and Likens, 1991). Sediment yield from the 1-km² catchment of Mirror Lake within the Hubbard Brook watershed is similar at around 2 t km⁻² yr⁻¹. Both of these values are close to the average Holocene sediment yield derived from Mirror Lake sedimentation studies of 8.6 t km⁻² yr⁻¹, after subtracting out the high (6 t km⁻² yr⁻¹) sedimentary diatom silica component in the lake (Davis et al., 1985). Low sediment yield at Hubbard Brook is primarily attributed to highly porous soil, low frequency of overland flow and the coarse texture of channel and channel bank sediment. The other lake sediment-based sediment yield study (cited earlier in chapter 2) is by Davis (1976) for Frain's Lake in southern Michigan. Average

Holocene sediment yield under fully forested conditions was calculated by Davis to be about $9 \text{ t km}^{-2} \text{ yr}^{-1}$. This value is near that from Thompson Lake despite the fact that Frain's Lake is a completely closed basin with mineral sediment input resulting solely from sheetwash off adjacent slopes, and is thus fundamentally different in erosion and transport mechanisms than Thompson Lake. Post-settlement sediment yields at Frain's Lake were computed to be much higher than Holocene rates, at $90 \text{ t km}^{-2} \text{ yr}^{-1}$.

On the west coast, at the H. J. Andrews Experimental Forest in the Oregon Cascade Mountains, sediment yield over eight years from the fully forested control watershed is similar to that at Coweeta at $3.5 \text{ t km}^{-2} \text{ yr}^{-1}$ (Frederikson, 1970), a value less than average yield at Thompson Lake. This relatively low value occurs despite the high relief of the catchment and nearly twice as much precipitation as received at Thompson Lake. Undisturbed soils on the volcanic (tuff, basaltic/andesitic breccia) bedrock are porous and resistant to topsoil erosion. Disturbance leading to removal or compaction of these soils has been minimal. Mass failure is the most important erosion process in the basin (Frederikson, 1970), therefore, periods of higher sediment yield would likely be tied to episodic occurrences of these events. Other differences such as the amount of precipitation received as snowfall, the resulting seasonality of runoff, and the type of vegetation present (coniferous vs. deciduous) may determine the low sediment yield of this site.

Controls of Sediment Yield

Assuming that Figure 14b represents an accurate portrayal of changes in sediment yield over the last 29 years, how can these changes be explained? Chapter 2 dealt with several

factors capable of producing changes in sediment yield through time. These include: 1.) contemporaneous land use changes leading to changing erosion rates in the catchment, 2.) change in sediment delivery ratio (SDR) in response to medium-term effects of logging, including the dynamics of coarse woody debris in the channel, and 3.) change in precipitation in terms of annual totals, seasonal distribution, frequency-magnitude characteristics, or a combination of these.

Over the last 29 years, the Thompson Lake catchment has been recovering from timber cuts earlier in the 20th century, cultivation of orchards on slopes directly marginal to the lake itself, and more recently, selective and clearcut logging on small areas in mid- and high-elevation portions of the catchment. Changes in catchment characteristics over the study period, then, are mainly forest regrowth after selective cutting in the 1950s and 1960s and after one small patch clearcut shown on air photos in 1990 in the northern headwater, and continuing timber maturation throughout the catchment. Selective timbering in the late 1950s and early 1960s may have increased the total amount of discharge and perhaps the variability in magnitude and frequency of flood events. Hillslope sediment production associated with selective logging may have been high close to logging roads visible in air photos. The isolation of logging sites and roads with respect to stream channels and the wide flat stream marginal topography of the second-order portion of the catchment would have minimized the movement of any hillslope sediment into this channel segment. But logging also occurred along both first-order streams at mid-elevation in the catchment, possibly contributing more sediment directly to the channel there. As a result, suspended sediment loads may have been higher during the 1960s. Also, increased discharge could have increased flow capacity and competence, remobilizing stored sediment within the channel.

Under these conditions, gradual recovery from logging over the last 29 years might be expected to reduce through time the sediment yield from the catchment. The sediment yield diagram (Fig. 14b) shows relatively high sediment yield for the first 15 years of record, as well as high sediment yield from 1992 to 1995, the most recent period. This pattern is not one of gradual decline in sediment yield that might result under conditions of progressive forest recovery. It is possible that a gradual decline early in the record (subzones 1a and 1b) has been obscured by the effects of the Agnes event. The destabilization of stored sediment in the channel, as well as of the channel banks themselves by the storm might be responsible for high subzone 1b yield by way of increasing availability of easily eroded sediment, increasing yield through time. A similar response to a high-magnitude storm in the Lainbach River basin, Bavaria, has been noted by Gintz et al. (1996). However, this phenomenon does not explain the low sediment yield of zone 2 and subsequent rise in sediment yield apparent during the last few years of record. High initial (subzone 1a) values of sediment yield may be associated with high initial rates of lake bank retreat that have declined through time, rather than with upland logging disturbance.

Changes in the size and frequency of coarse woody debris (CWD) dams during forest regrowth and maturation affect sediment yield by influencing in-channel sediment storage. Logging removes the source of larger tree boles and branches that continuously supply CWD to the stream channel as older debris dams decay and fail through time. Forest regrowth provides abundant small woody debris as forests thin and small early-successional species die off, but this debris cannot replace large, long-lived debris dams. The persistent effect of logging in terms of CWD input could be an influence at Thompson Lake. Today, CWD dams are rare at this site relative to their abundance in other less disturbed Appalachian catchments. In undisturbed portions of the Coweeta

watershed in North Carolina and at Hubbard Brook in New Hampshire, there are 20 to 30 debris dams per 100 meters (Webster et al., 1988; Likens and Bilby, 1982). At Thompson Lake there are only 2.7 debris dams per 100 meters. Thompson Lake debris dams involving woody debris greater than 10 centimeters in diameter most often are at an advanced state of decay; this is especially true for the few debris dams represented by individual logs of large diameter. In contrast, abundant finer CWD is much less degraded, indicating that it continues to be generated within the catchment. Given the small number of CWD dams at Thompson Lake, it seems likely that their destruction without replacement has occurred for longer than the last 29 years, as a result of multiple earlier episodes of timber harvest within the catchment.

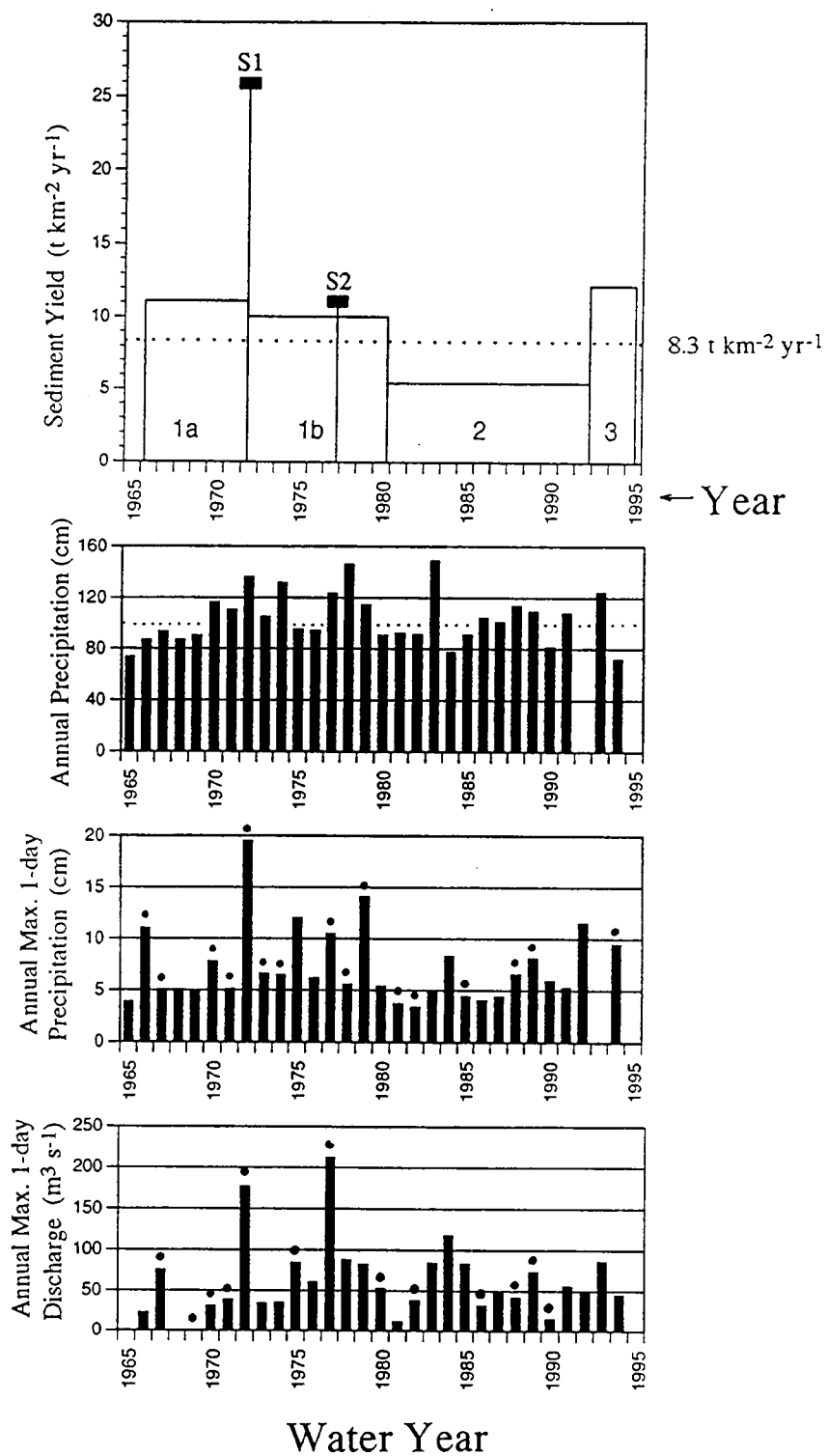
The trend in sediment yield expected from progressive net loss of CWD dams through time would be one of increasing sediment yield as stored sediment is released from failing debris dams. Such a simple pattern is not in evidence at Thompson Lake, therefore, as a lone mechanism explaining the sediment yield pattern, debris dam loss is inadequate. When coupled with hydrologic changes associated with logging, a more complex pattern may result. For example, higher runoff, flow capacity and competence, and stream margin disturbance resulting from tree removal in the late 1950s and early 1960s could account for higher sediment yield early in the record. Decreased capacity of flow later after forest regrowth (e.g., zone 2 low sediment yield) could be later countered by increased in-channel sediment availability and sediment delivery ratio as CWD dams begin to fail. Testing of such a scenario would require far more detailed process information and much greater time resolution of sediment yield than was produced by the current stratigraphic study.

There is better evidence available to link the reconstructed sediment yield pattern to patterns of precipitation over the 29 years since reservoir construction. Precipitation

records from Mt. Weather (and surrounding stations) show that the pattern of both total annual precipitation and annual maxima in one-day precipitation since 1965 generally corresponds to changes in sediment yield at Thompson Lake (Fig. 15). During the late 1960s, total annual precipitation was lower than the 100 cm average. Higher than average precipitation occurred throughout the mid- and later 1970s, precipitation was lower during the early 1980s, higher in the late 1980s, and variable in the early 1990s. High annual maxima in one-day precipitation were rare from 1965 to 1972, relatively common from 1972 to 1980 (including peaks from Hurricane Agnes and other large storms), less common during the 1980s, and again common in the early 1990s. The record of maximum daily average discharge at Goose Creek shows a pattern similar to that of precipitation (Fig. 15), with low annual discharge maxima occurring during the 1980s, except for maxima during the years 1983, 1984 and 1985. These trends mirror the trends in sediment yield with the exception of initially high sediment yield of zone 1a. This high sediment yield may be a result of higher lake bank erosion immediately following impoundment of streamflow.

The earlier (monitoring) observation that suspended sediment concentration is related to the occurrence of precipitation rather than directly to discharge implies that the precipitation record should be a better predictor of sediment yield. Control of sediment yield by total precipitation and frequency-magnitude characteristics of precipitation events provides a simpler explanation for the lake sediment record than explanations based on contemporaneous land disturbance. Similarly, Dearing and Flower (1982) found that lake sediment magnetism was strongly related to rainfall at a site in northern Ireland, and Dearing and Foster (1986b) found a strong relationship between sediment yield and annual precipitation at a lake catchment in Midland England. Accepting this relationship

FIGURE 15. Zonal average sediment yield vs. precipitation and discharge data from Mt. Weather and Goose Creek. Dots above histogram bars indicate event occurrence in May through November.



requires that the precipitation record be reflected in magnetic susceptibility profiles, since these reflect particle size and hence discharge and precipitation controls.

Comparison of these two records gives a counterintuitive result: with the exception of proposed storm event strata, lower susceptibility occurs during wetter stormier periods, and high susceptibility occurs in many cores during the relatively dry 1980s. In particular, the precipitation record must be reconciled with three characteristics of zone 2: 1.) Low sediment yield, 2.) High magnetic susceptibility and coarse sediment dominance in zone 2 from many cores, and 3.) High sedimentation rate predicted by the Pb-210 model at grid site A3, despite low sediment yield. I propose an explanation for these observations based on the occurrence of summer drought and otherwise low precipitation totals common during the 1980s. These conditions would have resulted in lower pool elevations in the reservoir, especially during abnormally strong summer drought periods that occurred most notably in water years 1984 and 1990 and to a lesser extent in 1980, 1981, 1982 and 1985. Lower pool elevations relative to those during the 1970s would have created shallower conditions north of the island where the reservoir bed drops off very gradually toward the deep-water zone. Thick zone 1 accumulations at core D1 (Fig. 10) indicate that this area had consistently received sediment from the lake inlet up until and probably through zone 2, as well as during recent times.

Under shallow flow conditions, a larger portion of total load and the coarse fraction would have been more easily transported to deeper water during infrequent elevated flows. Because mass-specific susceptibility was found to be more strongly related to the amount of fine sand and coarse silt rather than coarse sand fraction, catastrophic floods would not be necessary to produce the marked shift in susceptibility which occurs at zone 2. I propose that infrequent moderate flood events were sufficient to

transport coarse silt and fine sand to deeper water during low pool conditions. That moderate floods are acting together with potentially low pool elevations to produce the observed sediment yields is suggested by the very rapid rise in susceptibility at the beginning of zone 2 and the dual-peak nature of the zone 2 susceptibility profile displayed in cores C2 and A3 (Fig. 10). In these cores, as well as cores C1, C4 and D1, D3, the initial peak, which often occurs as a disturbance on the rising side of broad deltaic zone 2 peaks, appears less pronounced than the succeeding peak dominating the zone. This succeeding peak could represent the major 1989 dendrohydrological event.

The path of coarse sediment movement is reflected in the locations of cores for which zone 2 has high susceptibility (Fig. 8a). Highest susceptibility occurs in zone 2 of core D1; moving to deep water, high zone 2 susceptibility follows cores C2, B2, and A2 and A3; this is the line of deepest water extending from north of the island to the impounding structure (Fig. 3b). High X^{lf} of zone 2 in deltaic cores south of the island would represent coarse sediment accumulations associated with moderate floods as in the case for zone 2 north of the island. Lower zone 2 X^{lf} values displayed in other cores from this south side of the island and on the south side of the lake in general may reflect differences in volume of flood flow north and south of the island. Alternatively, there could have been resuspension or "winnowing erosion" of older lake sediments in the shallow area north of the island, and their redeposition down-flow in deeper water. This process is suggested by thinner total accumulations of core E1.5 up-flow of the core (D1) containing maximum zone 2 X^{lf} values, possibly associated with redeposition of coarse silt. Increasing the proportion of sediment load deposited in deep water would explain the spatial homogeneity of the zone 2 isopach map (Fig. 11b) relative to maps for zones 1 and zone 3. Further evidence of such a process may be seen in Cs-137 profiles of zone 2 from cores A3 and C1 (Fig. 12), both of which show a rise in Cs-137 concentrations

during a time of general Cs-137 decrease. Cesium fallout has been virtually zero since the early 1970s, and a catchment erosion source is unlikely during this time period of low magnitude flows and unlikely erosion of Cs-137-enriched topsoil.

A difficulty with this hypothesis is that it implies a redeposition of adsorbed Pb-210 in deep water as well. High sedimentation rates in core A3 indicated by the CRS model for zone 2 should be associated with dilution of fallout by increased input of Pb-210-poor sediment rather than redeposition of Pb-210-rich sediment. This discrepancy suggests that a significant proportion of total sediment accumulation at core A3 during zone 2 is derived from catchment erosion; this is a likely scenario in any event, especially under low pool conditions. Also, since delta deposits that could have been later eroded would have already been sorted, only relatively coarse (with regard to Pb-210 and Cs-137 adsorption) sediments might have been available for mobilization. If so, most sediment from this source may have traveled only a short distance, for example, more strongly influencing transects D and C. Supporting this idea is the much more pronounced zone 2 Cs-137 peak in core C1 (a possible redeposition effect), as compared to that in core A3. If so, the increased sedimentation rates in core A3 (deep water) would simply reflect the relative lack of deposition in the deltaic area under low pool conditions.

My interpretation of the environmental history at Thompson Lake is that the dominant control of medium-term sediment yield has been precipitation. Periods of high annual precipitation totals, and with more frequent moderate- and high-magnitude flooding, are associated with increases in sediment yield. This was the case at Thompson Lake through most of the 1970s and the early 1990s. Lower precipitation and discharge occurred through most of the 1980s and produced lower sediment yield during this time. Increased values of X^{lf} during the 1980s (zone 2) are associated with a change in the location of the principal sediment accumulation areas brought about by lower pool

elevations in the reservoir. A return to wetter conditions in the 1990s may have reversed this trend. It is likely that changes in hydrology brought about by selective logging prior to dam construction, and changing in-channel sediment storage associated with failing debris dams have had some influence on medium-term sediment yield. Neither of these influences however, appears more important than precipitation as a control of sediment yield at Thompson Lake.

Frequency-Magnitude Analysis

Frequency-magnitude analysis requires that sediment yields from individual storm events be quantified. Wolman and Miller (1960) and Leopold et al. (1964) have presented examples of larger streams for which infrequent high magnitude events have made little contribution to long-term sediment yield (see chapter 2). For these larger streams, events with return times less than five years accounted for 90% of sediment yield based on continuous monitoring data. The evidence for this general relationship for large streams in the humid zone seems reasonable (Dury, 1980). As basin size decreases however, the percentage contribution of infrequent high-magnitude events has been hypothesized to increase due to the greater temporal variation in flow (Leopold et al., 1964). Velbel (1985) has concluded from different evidence that long-term physical denudation in small Coweeta watersheds must be dominated by low-frequency, high-magnitude events. Velbel argues that 50 to 95% of physical erosion ("landslides" and "severe storms") must occur by large events producing sediment yield much greater than the measured value of $3 \text{ t km}^{-2} \text{ yr}^{-1}$ to account for the presumed long-term dynamic equilibrium of the Appalachian Mountains.

Whether or not this pattern occurs, the degree to which it occurs, and how this varies from place to place are poorly understood geographical questions that would require enormous time and expense to answer using traditional direct monitoring in small catchments. To address these questions using lake sediment-based sediment yields requires that sediment inputs from individual storms be both identifiable and quantifiable in the sediment record. The identification of storm strata in this project relies on the relationship between particle size and X^{lf} . High competence of flood flows results in coarse particle strata with sharp, high X^{lf} peaks over lower, relatively uniform background values. Two coarse-particle strata associated with high magnitude events (S1 and S2) have been identified in zone 1, traced from core to core across the lake, and quantified in terms of volume and mineral mass. For event S1, the up-core decline in X^{lf} from the peak varies from an immediate (within 1-cm depth increment) return to low conditions (e.g., cores A3 and C0.5; common in deep-water and middle-lake cores), to steady linear decline over a few centimeters (e.g., core H3.8) in thicker sequences (Fig. 10). Large change in X^{lf} over one centimeter suggests that bioturbation is ineffective in terms of sediment homogenization. Where such a X^{lf} profile exists, it is likely that any associated storm sedimentation is contained within the peak area defined by maximum X^{lf} ; that is, storm sedimentation is less than one-centimeter-thick within a X^{lf} peak defined by a one-point maximum with steep initial rise and steep subsequent decline to pre-event values. Where gradual decline occurs as in the case of event S1 in core E3.2, and even more pronounced in core E3.8 event S2, two possible explanations exist. Gradual X^{lf} decline may be caused by grading associated with differential settling of sediment from a single storm event of first coarse particles of high X^{lf} , and successively finer sediments of lower X^{lf} . The second possibility is that a coarse sediment layer of

very high X^{lf} is mixed through bioturbation evenly up-column, with benthic fauna unable to completely penetrate through the base of the layer.

Neither of the two explanations alone would account for gradual linear decline in X^{lf} like that for event S2 in cores E3.2 and E3.8. Both explanations are compatible with the form of most deep-water profiles. I used this reasoning in assigning boundaries to event layers S1 and S2 (Fig. 10). For event X^{lf} peaks represented by a single high X^{lf} value, thickness of the storm sediment layer was assumed to be one centimeter. For such peaks, which are most common in deep-water cores, sediment thickness is probably overestimated, because the chances of selecting the exact bounds of a storm layer during core subsampling at 1-cm intervals is low. Most broad X^{lf} peaks begin, as do narrow ones, with a sharp rise over a 1-cm increment, making the selection of the lower boundary simple. Upper boundaries were placed in the middle of subsequent declines at the level at which peak height is reduced in half, or at which steep decline occurs. The choice of this position is based on the foregoing consideration of particle settling and bioturbation.

Because of the overestimation of event layer thickness in deep-water cores, sediment yield estimates for these events represent upper limits. Sediment mass from event S2 corrected for trap efficiency is 100.6 tons, equivalent to 10.9% of the total estimated 29-year catchment sediment loss (921.8 tons after trap efficiency, lake bank erosion and airborne input corrections). Corrected sediment mass from event S2 is 45.7 tons, or 4.9% of the 29-year sediment yield estimate. Precipitation from Hurricane Agnes in this area had a recurrence interval of greater than 100 years (Bailey et al., 1975). Extrapolating the 29-year average sediment yield over this time period (3,179 tons) and recalculating the percentage sediment contribution for event S1 gives 3.2%. This result means that 96.8% of the 100-year average sediment yield is associated with moderate and

small events even in this small headwater catchment. This high percentage is only slightly lower than that measured in much larger Potomac River sub-catchments (Table 5) and therefore does not support the hypothesis that the percentage contribution from large storms increases proportionately with basin area decrease. Because I am unable to establish the precipitation event responsible for event S2 sedimentation, and therefore obtain its recurrence interval, a similar calculation cannot be made for event S2. The fact that two apparently low-frequency, high-magnitude storms occurred at this site in the same decade highlights the difficulty of using recurrence intervals to predict sediment yields.

The above extrapolation of medium-term sediment yield over 100 years for the purpose of incorporating recurrence interval into the importance value of Hurricane Agnes is difficult to justify on either empirical or theoretical grounds. The concepts of frequency and magnitude are fundamentally grounded on the idea of stationarity, a term meaning that the mean and variance of an event sequence do not change through time (Thorn, 1988). Event sequences occurring over a short time span, for example, one corresponding to the "steady time" ($0-10^{-1}$ years) of Chorley and Kennedy (1971), may reflect stationarity. Short-term monitoring programs reflect this timescale. Over longer periods (>10 years; "graded" and "cyclic" time), however, including time periods normally covered by lake sediment analyses, stationarity does not hold. Church (1980) has discussed three particular problems of extrapolation over these timescales: 1.) trend (change in mean value), 2.) persistence (short-term serial dependence, or autocorrelation), and 3.) intermittency (long-term serial dependence). The existence of these effects means a lack of stationarity. For example, Dury (1980) describes a "step function" shift in climate and associated sediment yields in the mid-latitude northern hemisphere beginning about 1950. Increased variability in weather and climate after this

TABLE 5. Medium-Term Importance of Event S1 (Hurricane Agnes) Sediment Yield for Selected Potomac River Sub-Catchments.

Sub-Catchment	Catchment Area (km ²)	% 29 year Sed. Yield	% 100 year Sed. Yield
Crooked Run*	3.6	10.9	3.2
Anacostia	54.6	3.2	0.9
Monocacy	2,116	5.8	1.7
Potomac, Pt. of Rocks	24,996	3.8	1.2

*This study.

shift means that frequency-magnitude characteristics of discharge since 1950 may be different from those during the period 1900 to 1950 (Dury, 1980). If this shift is applicable at Thompson Lake, the 100-year extrapolation could be in error because it spans these two periods of different frequency-magnitude characteristics. In addition, it is incorrect to equate climatic events with geomorphic events. Even if climatic events are independent of one another, it does not follow that geomorphic events are independent as well (Kelsey, 1982).

The theoretical basis for predicting greater influence of high-magnitude, low-frequency events in small basins remains valid, and the results of this study do, to a degree, conform to this expectation. The changes at Thompson Lake are only weakly proportional to changes in basin area, suggesting that some spatial variable other than basin size may act as a significant influence on the spatial variability of sediment yield frequency and magnitude. I speculate that in the case of Thompson Lake, this variable is actually the combination of surficial geology and relief characteristics, both of which are linked to morphologic evolution of the catchment during the Quaternary Period. Thick, weathered colluvium residing in stream valleys, and particularly notable along the second-order stream segment, is deeply incised by the second-order stream, and provides abundant erodible sediment. This results in moderate sediment loads even during relatively small but frequent precipitation events. That is, sediment yield is not supply limited; major storms are not necessary to introduce sediment into streams. Also the wide, gentle surface of the valley fill disconnects steep mountain slopes from the main channel, reducing any hillslope colluvial input during larger storms.

Error

A number of error sources have already been evaluated in preceding chapters. These include error associated with spatial interpolation of sediment thickness, the indeterminate location of the zero-centimeter isopach, errors in estimated quantities like average runoff and trap efficiency, and error resulting from using precipitation data from a station nearby, but not actually within the study catchment. These errors primarily affect the absolute values of sediment yield estimates. There are two main error types that affect the relative values of sediment yield between zones and between storm sedimentation events, and therefore also interpretations of sediment yield controls and frequency-magnitude characteristics. The first has to do with the sensitivity of zonal and event sediment mass calculations to the placement of zonal and event stratigraphic boundaries. Assuming the zonal correlation scheme is correct and that the rules for establishing boundaries are used consistently from core to core, this type of error should be held at a minimum. If the correlation scheme were incorrect, a large amount of error would be involved. The marked difference between deltaic and deep-water X^{lf} profiles is the main obstacle to overcome. Although proving the correctness of a particular correlation scheme linking these two areas is not possible, I believe the proposed scheme best relates broad similarities in positive X^{lf} fluctuations observed in cores away from the lake margin. Relying solely on deep-water cores (below four meters water depth) which are reliably correlated, calculations of sediment yield for each zone gives a temporal trend subdued in contrast, but otherwise similar to that derived from the whole lake: high during the late 1960s and the 1970s, lowest values in the 1980s, and high values in the early 1990s. Also, storm events S1 and S2 comprise 7.4% and 3.3% respectively of the 29-year sediment yield, and event S1 sediment yield equates to 2.1% of the 100 year

extrapolation of this estimate. Also the relative importance of events S1 and S2 sediment contributions ($S1:S2 = 2.2$) is the same in deep water as in shallow water. These results confirm that a certain amount of proportionality is maintained over time between deposition at the delta and in deep water, and also support the proposed core correlation scheme.

The second potential error of this type pertains to assumptions for use of the constant rate of supply (CRS) model for Pb-210 accumulation and the resulting accuracy of the depth-age curve used to date zonal boundaries and storm events. The CRS model is based on 1.) the continuous fallout of unsupported Pb-210 onto the lake and its incorporation into sediment in the water column, and 2.) the variable dilution of fallout by input of Pb-210-poor fluvial sediment from the catchment. The first assumption is valid at most sites; Pb-210 does fall out onto areas continuously, although at a rate that varies somewhat from year to year, and Pb-210 is rapidly scavenged in the water column by mineral and organic sediment particles and incorporated into lake sediment. The second assumption is more problematic. Like Cs-137, Pb-210 accumulates in topsoil which may be eroded and enter the fluvial system. Unlike Cs-137 today, which has undergone substantial decay and erosional loss since its deposition, Pb-210 may be present in channel sediment in temporary active storage at near or greater than its current fallout level. At Thompson Lake, the largest sediment source per unit basin area is channel bank subsoil, but sediment input from other sources is suggested by the monitoring data. Topsoil inputs are of lesser importance due to the disconnection of hillslopes from the channel, associated with wide valley fill. However, during large storms, more hillslope topsoil may enter the channel, and more in-channel storage may be mobilized, both leading to greater than normal inputs of Pb-210 from the catchment, and a resulting lack of dilution of Pb-210 fallout in the lake. The unsupported Pb-210 profile

from core A3 suggests that this could be happening at Thompson Lake. Pb-210 values (per gram) in this core sometimes fluctuate directly with X^{lf} values in the lower part of the core, coincident with coarse-particle storm sediment layers. The effect of these inputs would be to reduce calculated sedimentation rates and overestimate the ages of affected zones. If this is the case for Thompson Lake at zone 1, in which susceptibility fluctuations and the precipitation record indicate at least two large storm events, sediment yield in this zone is underestimated, and it is overestimated for succeeding zones. Note that this circumstance would intensify the differences in the sediment yield diagram (Fig. 14) rather than subdue or reverse them. Also, the ratio of storm sediment mass to average medium-term sediment yield would not be changed by altering the ages of zonal boundaries.

The CRS model assumptions appear to be met most of the time at Thompson Lake. The likelihood that this is not always so introduces a problem into the accuracy of Pb-210 depth-age curves. Without a more detailed understanding of transport and sinks of Pb-210 at the site, it is difficult to further evaluate the accuracy of the depth-age curve. Fortunately, such a quantitative evaluation may not be critically important, since it would not affect frequency-magnitude analysis, and since the qualitative evaluation appears to support the temporal pattern of average sediment yield proposed earlier.

IX. CONCLUSIONS

In this study I have brought together several means of indirect monitoring to reconstruct past changes in sediment yield from a small lake catchment. The 29-year record from Thompson Lake is shorter than most preceding stratigraphic records from other reservoirs, the majority of which are located in the United Kingdom. Because of the record's brevity, major transitions in sediment yield resulting from contemporaneous and substantive land use changes are not sampled. Longer records are needed to understand the effects of various land use types on average sediment yields, and the persistent influence of a given land use type on future sediment yields at a site. In the US, few studies of sediment yield have involved detailed analysis of lake sediment stratigraphy, and few of these come from sites that are similar in morphology and sediment delivery processes to the majority of monitored catchments. This dissimilarity makes comparison of sediment yields more difficult. The usefulness of the results obtained in this study are enhanced by the site's similarity to the majority of monitored catchments. The 29-year length of record also represents an improvement over the short-term (or complete lack of) records typical of monitoring studies in small catchments. However, the longer record should not necessarily be considered superior to short-term monitoring records, but instead should be viewed as offering a different and useful perspective not easily obtained by monitoring. The two different types of record address different time scales of process operation.

Results from Thompson Lake contribute to methodological, empirical and theoretical aspects of watershed dynamics. This work represents the first time a lake sediment-based sediment yield reconstruction has been applied to a reservoir in the eastern US in which fluvial discharge is the primary mechanism of sediment input. It is

the first in the eastern US relying on magnetic susceptibility for reservoir sediment stratigraphy and as an indicator of other sediment dynamics in small watersheds. It is also the first to incorporate dendrohydrological data as a means of understanding the sediment yield history preserved in lake sediments. The main goals were to determine average sediment yield since reservoir construction, as well as changes in the average over time. As part of the latter goal, another objective was to derive information on the frequency and magnitude of sediment yield at this small areal scale, that is essentially unavailable from direct monitoring programs. An additional purpose was to determine controls of sediment yield over medium-term timescales especially with regard to contemporaneous land use and ecological change. Although it is clear from the prior chapters that many ambiguities result from the lack of direct monitoring data, all of these goals and purposes were at least partially realized.

I derived an estimate of average 29-year sediment yield that is consistent with values predicted for fully-forested conditions in this and some other eastern US areas. Variations in X_{lf} through time correlative across the lake allowed the resolution of the sediment yield record into smaller time periods. Sediment yields for these periods appear to correspond with changes in annual precipitation totals and other precipitation characteristics in the region. Another distinguishing feature of this study was the identification and quantification of sediment yield from at least one individual high magnitude stream discharge event in the lake sediment record, an achievement not duplicated in other lake sediment studies. Sediment yield from this event, in proportion to the 29-year average yield, was found to be only slightly more important than that often cited for larger rivers. This finding is counter to the hypothesis that high-magnitude storms increase proportionately in importance with regard to medium-term sediment yield as basin size decreases. Reneau and Dietrich (1991) obtained similar results for an area in

the Coast Range of Oregon, relying on monitoring data. This specific finding at Thompson Lake suggests that controls other than streamflow characteristics are responsible for a major component of spatial variation in sediment yield. The extension of lake sediment studies to other carefully chosen sites in different environments within the regions affected by Hurricane Agnes should be helpful in understanding these other controls.

The biggest problem in the lake sediment-based reconstruction of sediment yield is how to derive an accurate chronology for recent sedimentation. Most earlier studies have relied on a combination of Cs-137 and Pb-210 as I did at this site. The use of either isotope involves meeting a set of site requirements that are best matched at sites where either organic productivity is the dominant sediment input, or catchment sediment yield varies little over time. Lakes with the first characteristic are usually "special case" basins without a well developed stream network flowing in or out. With regard to the second characteristic, if catchment sediment yield is known to vary little through time, it might be simpler and more accurate to use extrapolations from short-term monitoring to get at moderate-event sediment yields, and to simply wait for an individual high-magnitude storm to monitor for comparison. Excluding the special case catchments noted above, sites with complacent sediment yield records are rare, because discharge, especially in small catchments, does fluctuate widely in response to precipitation inputs. The CRS model of Pb-210 accumulation was constructed for use in such catchments, and can be reliable when the relative importance of different sediment sources is known. Unfortunately, this information can be difficult to obtain, and if obtained, cannot always be assumed to be valid for earlier periods in the catchment's history. The coupling of sediment source "fingerprinting" techniques (Peart and Walling, 1986) with lake sediment-based sediment yield reconstruction offers potential for solving this problem.

Site selection is key to the success of using multiple indirect methods to reconstruct sediment yield. Reservoirs, which are common in a wide variety of environments and which are commonly located within well-defined stream networks, offer opportunities for understanding the spatial attributes of sediment yield. The need for more finely resolved sediment yield records from lakes narrows considerably the number of useful reservoirs and therefore the locations available for study. A more difficult problem in many areas is finding an otherwise acceptable site much older than Thompson Lake having undisturbed lake sediment. This has been an important constraint on site selection in this study. If suitable sites are too few in number, lake sediment-based methods of sediment yield research may ultimately be of little use for studying the spatial variation of frequency - magnitude characteristics. My initial reconnaissance of the Blue Ridge Mountains indicates that, at least for this province, perfectly suitable sites may be few. If so, increasing the pool of potential study sites will require relaxing the suitability requirements. For example, relaxing the accessibility requirement may increase the pool of useful sites, while sacrificing the logistical ease of fieldwork at more accessible sites. This particular trade-off is probably worth pursuing, however, because the quality of resulting record remains high. Other requirements, such as high trap efficiency and simple lake bathymetry, are more important to the sediment yield reconstruction, and are therefore less easy to sacrifice. Part of the difficulty of finding suitable sites is simply determining how well a given site meets selection requirements. For example, the task of learning the dates of construction and subsequent degree of sediment disturbance of sites that appear promising on maps, is difficult and time consuming. For purposes of understanding sediment disturbance, and the degree to which a sediment sequence contains useful stratigraphic markers, rapid magnetic scanning of whole cores in the field using a core scanning sensor may be a useful technique. If a number of suitable sites

cannot be found using these means, combining stratigraphic studies from select sites, and total sediment mass surveys from less ideal sites in surrounding areas may be the most useful research strategy.

Future research directions suggested by the current study include using more precise particle-size - X^{lf} relationships to further constrain storm sediment mass within a layer, obtaining accurate data on lake bank erosion inputs, providing alternative means of chronology in youthful sediments, the development of better models to deal with existing chronological indicators, and the application of existing models to estimate past discharge from precipitation input data.

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APPENDIX

APPENDIX

Pb-210 Chronology

Total Pb-210 in a sediment layer is comprised of two components: 1.) unsupported Pb-210 associated with atmospheric fallout produced by decay of Radon (gas) in the Uranium decay series chain, and 2.) supported Pb-210 produced by decay of Uranium series elements from in situ lake sediments. Because the fallout of unsupported Pb-210 ($t_{1/2} = 22.26$ yr) is approximately constant through time for a given area, its accumulation and subsequent decay in lake sediment can be used as a dating tool. The value for the unsupported component is calculated by measuring the total Pb-210 and correcting for the supported component. The supported component is measured by determining the concentration of Pb-214 ($t_{1/2} = 27$ min.) which decays to Pb-210 via the decay products Bi-214 ($t_{1/2} = 20$ min.) and Po-214 ($t_{1/2} = 2 \times 10^{-4}$ s) (or Tl-210; $t_{1/2} = 1.3$ min.). The short half lives for the decay series Pb-214 to Pb-210 mean that Pb-214 found in undisturbed sediment layers at all points below the sediment-water interface originates by decay of Ra-226 ($t_{1/2} = 1,622$ yr) within lake sediment rather than from Rn-222 in the atmosphere. Therefore, if it is in secular equilibrium with Ra-226, Pb-214 can be used as a measure of supported Pb-210 to which it quickly decays. Thompson Lake sediment contains quantities of Pb-214 too low to measure by gamma spectrometry of bulk sediment samples and therefore appears to have a negligible supported Pb-210 component.

Given a downcore series of unsupported Pb-210 values, various models can be applied to provide age estimates for member depths of the series. In this dissertation I use the constant rate of supply (CRS) model derived by Appleby and Oldfield (1978) to

estimate ages for lake sediment strata. As the name implies, this model assumes a constant rate of supply of unsupported Pb-210 to lake sediment. Unlike some other models (e.g. the constant initial concentration (CIC) model) the CRS model accounts for changes in input rates of sediment. It does this by assuming that sediment not derived from fallout (ie. stream sediment) is low in Pb-210 and thus produces variable dilution of unsupported Pb-210 concentrations.

In this (CRS) model, the age (t) of a horizon at depth x is given by:

$$(1) \quad t = 1/k \ln(A(0)/A(x))$$

where k = decay constant for Pb-210 (0.03114 yr^{-1}), $A(0)$ = total residual unsupported Pb-210 in the sediment column, and $A(x)$ = total residual unsupported Pb-210 below depth x in the sediment column. Implied but not usually stated in descriptions of this equation is the assumption that unsupported Pb-210 decreases to a zero value through decay at some depth in the sediment column. This requires that a sediment sequence be at least several times the half-life of Pb-210 in age; the limit of Pb-210 dating is commonly assumed to be around 150 years.

To derive a chronology for sediment at Thompson Lake (29 years old) requires that a correction be made for the "missing" Pb-210 inventory that would be present in older sediment. That is, an estimate of $A(0)$ must be made for a hypothetical core from Thompson Lake that is much older than and has the same 29-year residual unsupported Pb-210 inventory as the Pb-210 profile obtained from Thompson Lake core A3. To do this, I set $A(x)$ in equation 1 equal to $A(0)$ minus the 29-year inventory at the lake, set $t = 29$ years, and solved for $A(0)$:

$$29 \text{ y} = (1/0.03114 \text{ yr}^{-1}) \ln\{A(0)/(A(0)-32.111 \text{ pCi cm}^{-2})\}$$

Solving for A(0),

$$A(0) = A(0)e^{0.9031} - 32.111e^{0.9031}$$

$$A(0) = 54.000 \text{ pCi cm}^{-2}$$

Adding this value for A(0) to each measured value for the last 29 years gives values that can be added cumulatively to get A(x). With these values, Equation 1 can be applied to derive age estimates for the last 29 years of sedimentation; the resulting basal date for Thompson Lake cores is thus anchored to the known 29 year age.

The derivation of equation 1 also gives an equation for sedimentation rate (r) in units of g cm⁻²yr⁻¹:

$$(2) \quad r = kA(x)/C$$

where C = the concentration of Pb-210 in pCi g⁻¹ dry sediment.

VITA

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