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Key Points:
- Tropical Storm Irene-like flooding and erosion have occurred repeatedly
- Catastrophic erosion events result in persistent elevated sediment yields
- Climatically driven changes in twentieth century erosion exceeded human impacts

Supporting Information:
- Supporting Information S1
- Data Set S1

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Contrasting human versus climatic impacts on erosion

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Abstract Both human activity and climate change can influence erosion rates and initiate rapid landscape change. Understanding the relative impact of these factors is critical to managing the risks of extreme erosion related to flooding and landslide occurrence. Here we present a 2100 year record of sediment mass accumulation and inferred erosion based on lacustrine sediment cores from Amherst Lake, Vermont, USA. Using deposition from August 2011 Tropical Storm Irene as a modern analogue, we identified distinct event deposits indicative of destructive erosion events. These deposits record a prolonged (multidecadal) interval of enhanced erosion following the initial storm-induced landscape disturbance. The direct impact of human land cover alteration is minimal in comparison to the more recent twentieth century increase in the occurrence of catastrophic erosion linked to overall wetter conditions that favor high erosion rates and more easily trigger landslides during periods of extreme precipitation.

1. Introduction

Erosion, sediment production, and sediment dispersal are fundamental geologic processes related to landscape evolution, mass-transport between the continents and oceans, recycling of Earth materials, and soil availability. At the same time, these processes are directly related to the resource needs of humans, impact water quality, and ecological services, and when they occur rapidly during floods and landslides can have catastrophic impacts on humans and built infrastructure. The possibility of more frequent floods and landslides accompanying projections for warmer and wetter climatic conditions in the future [*Crozier*, 2010] underscores the need to understand the connection between climate and episodes of rapid erosion. However, the coincidence of widespread human alteration of the landscape [*Hooke et al*., 2012] with ongoing natural geomorphic processes and changes related to both natural climate variability and anthropogenic climate change makes it difficult to distinguish the relative importance of these various forcings over human time scales [*Huggel et al*., 2012; *Kasprak et al*., 2013]. Consequently, this study presents a continuous, 2100 year record of sediment accumulation in a natural in-stream lake (Amherst Lake, southeastern Vermont) as a means of quantifying long-term rates of sediment transfer from the surrounding watershed and identifying discrete, episodic depositional events resulting from floods and upstream landslides. Sediment deposition associated with Tropical Storm Irene (TSI), which caused widespread flooding and landslides in the northeastern United States in August 2011 [*Olson and Bent*, 2013; *Dethier et al*., 2015; *Yellen et al*., 2014], is used as a modern analogue for interpreting past depositional events. Periods of rapid sediment accumulation and inferred erosion are assessed in relation to a well-documented history of human land cover change, existing paleoclimate reconstructions, and historical flood records with the primary goal of quantifying human and climatic impacts on the magnitude and frequency of extreme erosional events.

2. Site Description and History

Amherst Lake (AL; 43.47°N, 72.70°W, 326 m elevation; Figure 1) is a relatively small (0.33 km^2), deep (~29 m) lake located within a glacially enlarged basin along the main stem of the Black River. The Black River (49.4 km^2 basin area) is the only source of surface inflow to AL, entering at the lake's north end and exiting to the south. The narrow confines of the valley stabilize the delta at the north end of the lake minimizing the influence of stream channel migration [cf. *Brown et al*., 2002; *Parris et al*., 2010]. The lake has a simple, elongated shape with a 29 m deep proximal basin separated from a distal 20 m deep basin by a 12 m deep sill. A high relief (675 m), high gradient watershed favors rapid routing of sediment [*Fryirs et al*., 2007], and an ample supply of fine-grained clastic sediment is provided by thick accumulations of unconsolidated glacial till and localized glacial-fluvial sediments that veneer the surrounding landscape [*Underwood and Springston*, 2014].

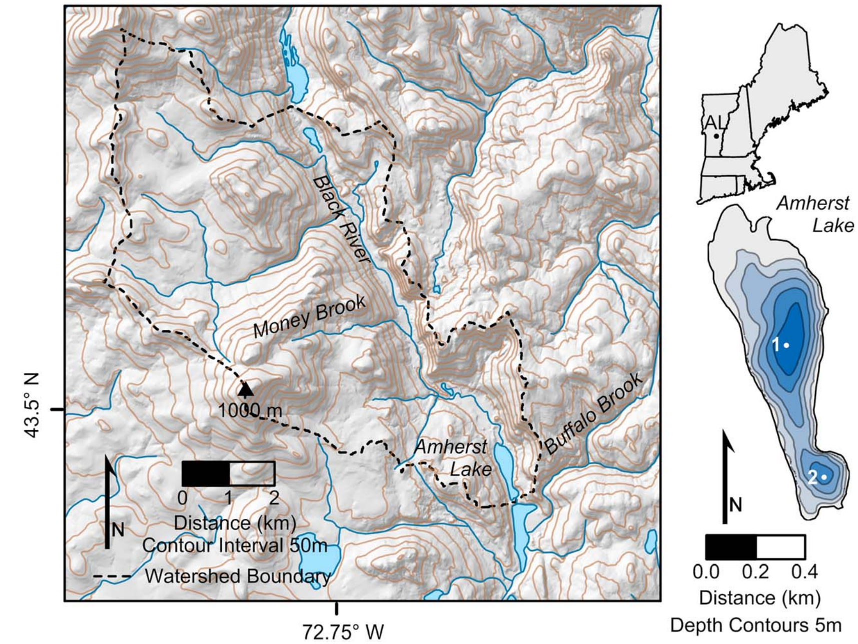

Figure 1. Watershed, bathymetric, and regional location map of Amherst Lake, Vt. Sites 1 and 2 coincide to cores P1 (43.48602°N, 72.70412°W) and D1 and D2 (43.48147°N, 72.70250°W), respectively.

The town of Plymouth, occupying much of AL's drainage basin. was chartered in 1761 (www.plymouthvermonthistoricalsociety.org). The population of Plymouth grew from 106 in 1790 to 500 by 1800. Following the local discovery of iron ore, population peaked at 1497 in the 1840s. A marble factory operated on the east side of AL in the 1830s, although agriculture and timber production dominated the economy in the 1800s. Though not part of the AL watershed, gold was discovered in nearby Buffalo Brook in 1857 and likely resulted in exploration in surrounding streams that drain into AL through the Black River. Mining operations were short-lived, and farms largely abandoned by the late nineteenth century. Historical photographs show a predominantly forested landscape surrounding AL by the beginning of the twentieth century (image LS00955 available from www.uvm.edu/landscape/). Today, the watershed remains dominantly forested with a small number of mostly seasonal residences scattered throughout the surrounding landscape.

Historical records document a history of significant flooding and landslide occurrence in the upper Black River watershed with known events in 1927 (the probable flood of record), 1936, 1938, 1952, 1973, 1976, and 2011 [*Underwood and Springston*, 2014]. Prior investigation of AL by *Noren et al.* [2002] identified prehistoric flood deposits in AL.

3. Methods

A gravity-percussion corer was used to collect surface-sediment cores from the proximal (154 cm long core P1) and distal (51 cm long core D1) basins of AL in August 2013 (Figure 1). A 4.5 m long, single-drive, piston-percussion core was collected from the distal basin (core D2) in March 2014 at the same location as core D1. All cores were split lengthwise and described prior to nondestructive analysis of magnetic susceptibility at 0.4 or 0.5 cm intervals using a Bartington MS2E surface sensor. An ITRAX scanning X-ray fluorescence core scanner [*Croudace et al.*, 2006] was used to obtain X-radiographs of split cores and determine continuous, down-core elemental abundances at 1.0 mm (P1 and D1) or 0.5 mm resolution (D2). Cores were scanned using a Mo X-ray source operating at 30 kV and 55 mA. Exposure times varied due to the need to balance instrument availability with the desire for longer count times and were 20 s (D2), 23 s (P1), and 30 s (D1). Percent loss on ignition (LOI) and dry bulk density were determined at 1 cm intervals following standard procedures [*Dean*, 1974]. Chronological control of recent sediments was determined by measuring the activity profile of the artificial radionuclide ^{137}Cs in core P1 using a Canberra GL2020R low-energy gamma detector and identifying the 1954 onset and 1963 peak activity horizons [*Appleby*, 2008]. Prehistoric age

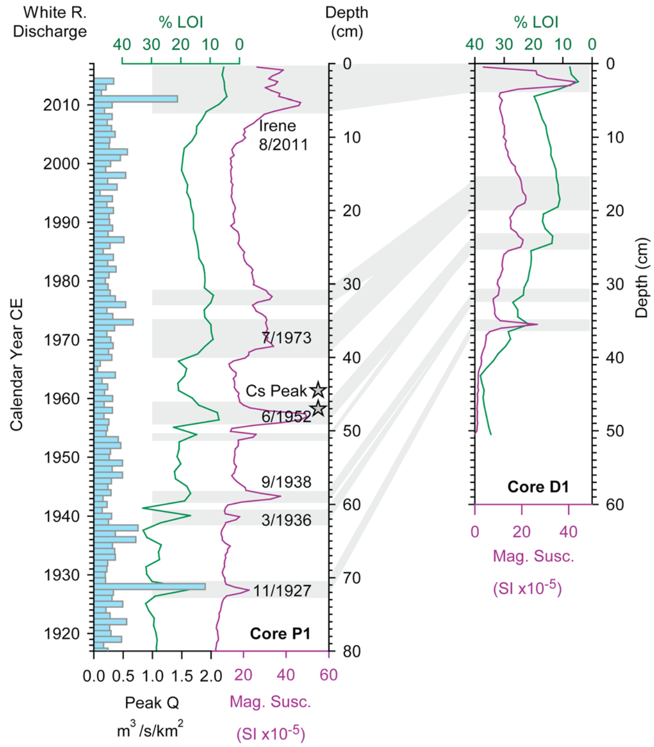

Figure 2. Down-core variations in LOI and magnetic susceptibility at proximal (core P1, only uppermost 80 cm are shown for clarity) and distal (core D1) ends of Amherst Lake compared to stream gage record of peak annual discharge from the White River. Note LOI scale is reversed. Event deposits (shaded gray) are correlative across the basin and coincide to known historical floods and landslide events in the upper Black River watershed and peak flow events in the White River as labeled. Onset and peak ^{137}Cs activity are indicated by stars.

control was determined by accelerator mass spectrometry radiocarbon dating of terrestrial macrofossils (leaves and pine needles). Radiocarbon ages were calibrated to calendar ages using the IntCal13 radiocarbon calibration curve [*Reimer et al.*, 2013] (Table S1 in the supporting information). Age-depth modeling of core P1 combined the ^{137}Cs stratigraphy with a single radiocarbon age using the CLAM routine and the weighted average of 1000 smoothed spline curves randomly fit through all possible ages based on their weighted probabilities [*Blaauw*, 2010]. Because core D2 lacked an intact sediment-water interface, a composite sequence for the distal basin was produced by splicing the upper 47 cm of core D1 to core D2 after identifying the overlapping portion of the cores. Age control for the distal composite record involved correlation of historical event beds to those identified in core P1 and an additional six radiocarbon ages modeled using the same procedure as for core P1. Although discharge of the Black River is recorded downstream of AL at North Springfield (USGS 01153000), the utility of this record is limited by flow regulation below AL after 1960. The nearest unregulated, long-term stream gage record to the study area is the White River gage at West Hartford (USGS 01144000), ~30 km from AL (Figure S1 in the supporting information). Based on the 31 year overlapping period of records prior to regulation of the Black River, 7 of the 10 largest peak annual flow events on the Black River are concomitant with peak flows recorded on the White River (Table S1). Consequently the longer, unregulated White River gage record was used to supplement historical accounts of floods and landslide activity in the Black River basin when evaluating sediment accumulation in AL in relation to historical floods.

4. Results

Cores P1 and D1 had intact sediment-water interfaces and preserved a distinctive tan, clastic (silt and clay) layer forming the surface deposit. This layer ranged from ~7 cm thick in proximal core P1 to ~4 cm thick in distal core D1 (Figure 2). Underlying sediment is dark brown, organic-rich gyttja occasionally punctuated by layers interpreted to be distinct event deposits similar to the surface deposit and consisting of sharp, conformable basal contacts followed by gradual transitions back to gyttja (Figure 2 and Figure S1). While fewer individual event deposits are identified at the distal site, the overall pattern of clastic layers is correlative between sites with a general thinning of individual layers and the overall sequence between the proximal and distal basins (Figure 2). Event deposits are anomalously dense, enriched in clastic material (low LOI and high magnetic susceptibility), and enriched in potassium (K) relative to background sediment (Figure S2 in the supporting information). ^{137}Cs first appears in core P1 at 47 cm and peaks at 44.5 cm. A calibrated radiocarbon age from a depth of 135.5 cm in P1 produces four distinct age ranges spanning 1640 C.E. through modern (Figure S2). While limited by increasing

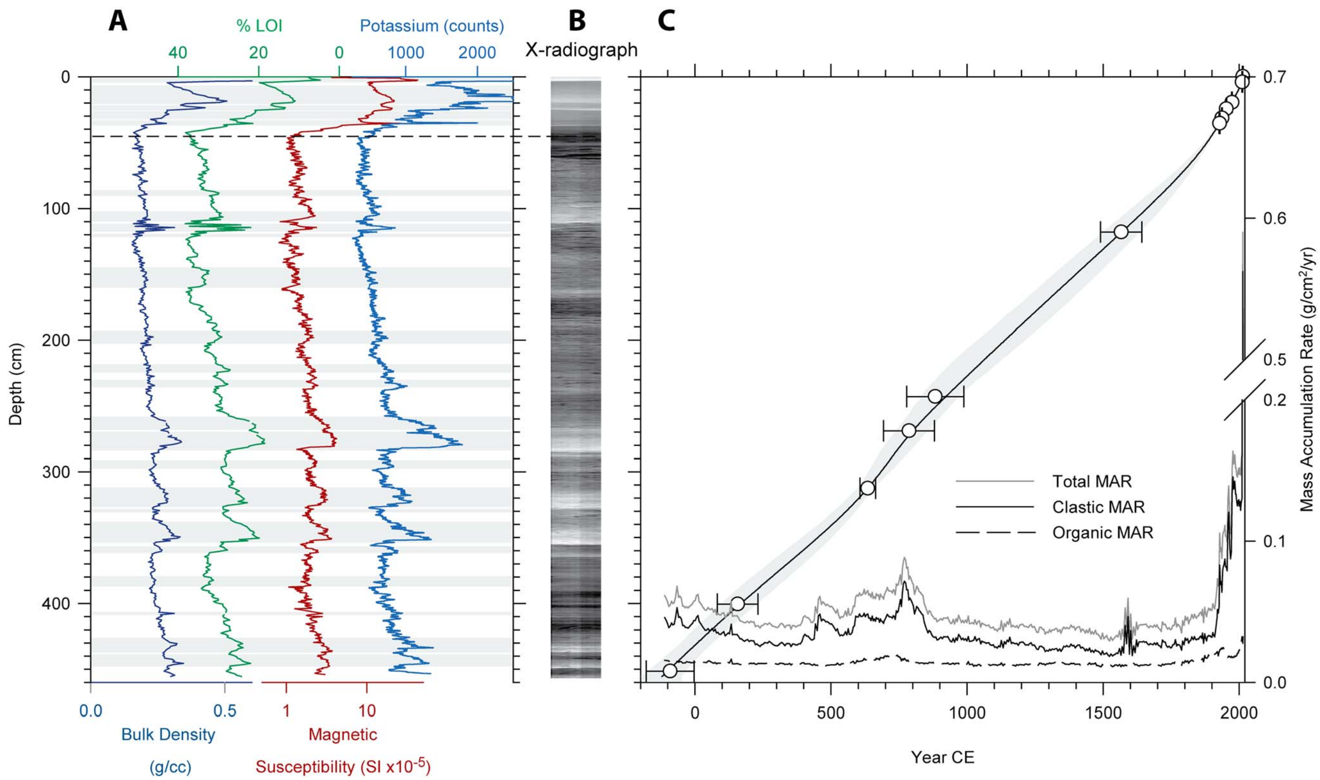

Figure 3. (a) Composite record based on cores D1 and D2 showing down-core variations in bulk density, LOI, magnetic susceptibility, and relative K abundance. Dashed line indicates the depth at which the composite record transitions from core D1 to D2. Event deposits are shaded gray. (b) X–radiograph of the distal composite record including cores D1 and D2. (c) Age-depth model of the distal composite record with 95% confidence intervals (in gray). Individual points represent median values of the full 2σ age ranges. Variations in organic, clastic, and total mass accumulation rate (MAR) through time are also presented.

uncertainty prior to the early twentieth century, the P1 age-depth model is sufficient for comparing the timing of event layers to known historical flood and landslide events in the Black River watershed and to peak annual discharge of the White River (Figure 2).

The clastic surface deposit found throughout AL was attributed to erosion from TSI in August 2011. K enrichment is a signature of TSI-sourced flood deposits throughout western New England attributed to mobilization of minimally weathered, K-enriched glacial deposits [*Yellen et al*., 2014]. The age-depth model for core P1 provides a 2σ age range of 1893 to 1932 for the event deposit at 72 cm depth (Figure S2). Alternatively, a linear trend extending from the surface through the ^{137}Cs age control points suggests an age for the same event layer between ~1925 and 1930. Correlation of this deposit with the flood of 1927 (the regional flood of record) aligns subsequent clastic layers with peak discharge events in the White River and all major episodes of destructive erosion documented in the upper Black River watershed since 1900 C.E. [*Underwood and Springston*, 2014] (Figure 2), providing a robust basis for interpreting prehistoric event deposits. While recording fewer events than the proximal site, the distal site remains sensitive to the most extreme historical events and benefits from being largely isolated from delta progradation, autocyclic depositional processes, and episodic slumping of the delta face. Consequently, the full 2100 year reconstruction relies on the composite sequence produced from the distal basin (Figure 3). Simultaneous occurrence of anomalously dense, low LOI, high magnetic susceptibility, and K-enriched intervals identify 24 event deposits with an average recurrence interval of ~92 years (Figure 3). The instantaneous mass accumulation rate (MAR) of bulk sediment was determined as the product of the linear accumulation rate defined by the age-depth model and the corresponding dry bulk density. Organic and inorganic (clastic) contributions to the total MAR were based on LOI (Figure 3). Variations in total MAR were dominated by changes in the clastic MAR, with periods of elevated MAR associated with more frequent event deposition. Elevated MAR occurs from −100 to 100 C.E., from 400 to 900 C.E., and after 1900 C.E. (Figure 3).

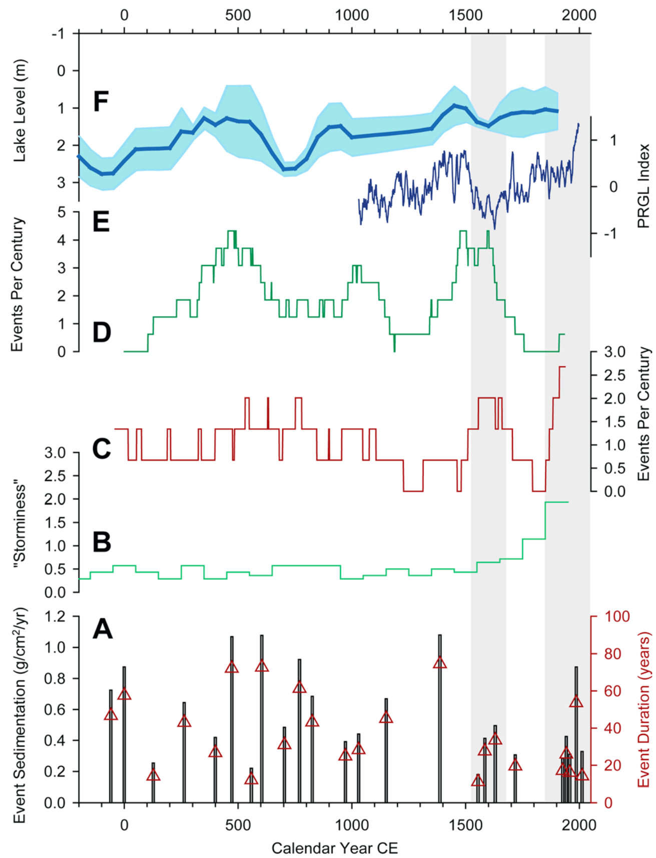

Figure 4. (a) Amherst Lake event deposits (bars indicate total mass accumulation and triangles event duration). (b) Northeast regional storminess [*Noren et al.*, 2002]. (c) Amherst Lake event frequency (events per century). (d) Salt Pond hurricane reconstruction [*Donnelly et al.*, 2015], (e) PRGL hydroclimate index with positive values indicating wetter conditions [*Hubeny et al.*, 2011], and (f) Deep Pond lake-level reconstruction [*Marsicek et al.*, 2013]. Gray shading indicates periods of more frequent event deposition at Amherst Lake.

5. Discussion

The lack of organic-rich sediment above the TSI deposit suggests that clastic sedimentation initiated by this event was ongoing at the time of core collection, 2 years after TSI. This observation is consistent with TSI-initiated landslides in the upper Black River watershed [*Underwood and Springston*, 2014] and throughout New England [*Dethier et al.*, 2015; *Yellen et al.*, 2014] having become persistent sources of sediment to streams. Event deposits throughout the record are characterized by an abrupt increase and then gradual decrease in proxy signals, suggesting that initial disturbances cause prolonged high sediment yields and corresponding deposition in AL. Based on the nearly constant organic MAR observed in AL (Figure 3), the duration of individual event deposits was estimated by dividing the total organic matter accumulated within each event layer by the long-term mean organic MAR (0.014 g/cm^2/yr). Resultant event durations range from ~10 to 80 years (Figure 4). Because this method does not take into account potential short-term variations in organic MAR associated with possible changes in productivity or due to input of terrestrial organic matter during floods, caution should be used when comparing the magnitude of individual events. Considered collectively, however, estimated event durations support the interpretation that event deposition is prolonged (potentially lasting decades) and results from persistent input of sediment related to the initial landscape disturbance. While events with shorter-duration impacts have been observed elsewhere in the region [*Rogers*, 2003; *Rogers et al.*, 2009], our results highlight the potential for long-term impacts from storm-related erosion and are relevant to the treatment of event deposits when modeling the age-depth relationship in sediment cores.

Variations in MAR at AL (Figure 3) and variations in the magnitude and frequency of event deposits (Figure 4) must be interpreted within the context of changing climate, possible changes in the magnitude and frequency of extreme floods, and human activity. Event magnitude, as defined by the total mass of sediment accumulated during an event (Figure 4), does not appear to be directly related to flood magnitude as recorded at the White River (Figure 2 and Table S2 in the supporting information). Analysis of historical climate data in relation to sediment yields initiated by the two largest flood events of the historical period, TSI and the 1927 flood, suggests that antecedent moisture conditions significantly influence the sediment yield produced by floods of a given magnitude [*Yellen et al*., 2015]. Thus, event sedimentation in AL should not be interpreted as an indication of peak discharge of past floods.

On longer time scales, the overall MAR in AL closely tracks changes in the frequency of event deposition (Figures 3 and 4), where event frequency is defined as the number of events per century occurring within a sliding 151 year window. The most significant change in AL MAR and inferred erosion is the rapid increase that begins in the twentieth century. Increased erosion and sediment flux are widely cited as a consequence of deforestation and agriculture activity [*Bierman et al*., 1997; *Costa*, 1975; *Hoffmann et al*., 2010; *Montgomery et al*., 2000; *Wilkinson and McElroy*, 2007]. The gradual increase in MAR from ~1750 C.E. to 1900 C.E. at AL likely reflects human activity and local land clearance that peaked in the mid-1800s. However, the rapid increase in MAR beginning in the twentieth century occurs after the watershed has largely reforested and hillslope erosion should have slowed. A delayed increase in MAR may result from disequilibrium between erosion and sediment yields [e.g., *Trimble*, 1977] and the remobilization of sediment temporarily stored as alluvium [*Phillips et al*., 2007; *Walter and Merritts*, 2008] by subsequent floods. However, AL is located near the high relief, high-gradient headwaters of the Black River drainage basin where temporary storage should be limited and downstream routing of sediment should be rapid [*Fryirs et al*., 2007]. The immediate depositional signal observed in AL in response to TSI and other documented landslides during the 1900s [*Underwood and Springston*, 2014] confirms rapid routing of sediment from hillslopes directly into the lake. Additionally, the regular occurrence of large-magnitude event deposits throughout the AL record indicates that twentieth century event deposits are not simply a consequence of increased sediment availability following eighteenth and nineteenth century human landscape disturbance. Instead, the twentieth century increase in MAR is associated with more frequent episodes of extreme erosion. While it is possible that human activity or built infrastructure could make the landscape more susceptible to erosion by floods and landslides, as mentioned previously, the AL watershed was reforested prior to twentieth century floods and remains largely undeveloped. Furthermore, landslides associated with TSI and other historical floods of the twentieth century occurred largely upstream from roads, bridges, and other infrastructure (Figure S3 in the supporting information). While the impact of land use change within the watershed cannot be completely ruled out, the lack of correlation between human disturbance and the pattern of erosion and deposition in AL suggests that human activity within the watershed is not the primary driver of changes in event deposition and MAR at AL.

Changes in AL MAR and event frequency independent of human activity imply that changes in erosion rates must be linked to climatic changes in the magnitude and frequency of storm events that initiate landslides in the watershed and/or that there is a natural mechanism influencing the sensitivity of the landscape to erosion in response to floods of a given magnitude. Existing reconstructions of past events likely to initiate destructive erosion around AL include compilations of past "storminess" based on the frequency of terrestrial in-wash layers in a network of regional lakes [*Noren et al*., 2002; *Parris et al*., 2010] and hurricanes [*Besonen et al*., 2008; *Donnelly et al*., 2015], which are a regionally significant source of extreme rainfall [*Barlow*, 2011]. The twentieth century increase in event frequency in AL (Figure 4) is consistent with increased storminess detected by *Noren et al*. [2002] and more frequent event deposits in AL centered around 1600 C.E. coincide with the interval of highest hurricane activity of the past 2000 years [*Donnelly et al*., 2015]. While this correlation suggests that AL event frequency and local erosion are sensitive to the occurrence of extreme rainfall events, limited agreement among the existing regional storm reconstructions precludes a definitive analysis. Alternatively, reconstructions of regional moisture balance [*Hubeny et al*., 2011; *Marsicek et al*., 2013] depict a trend toward increasingly wet conditions (Figure 4) that is consistent with tree ring reconstructions from New York, suggesting that modern conditions are the wettest of the past 500 years [*Pederson et al*., 2013] and long-term lake-level reconstructions that suggest modern conditions are among the wettest of the entire Holocene [*Shuman et al*., 2001]. Thus, the highest MAR in AL and most frequent event deposits coincide with the wettest period of the entire record. While event frequency and MAR do not consistently track regional moisture balance over the entire period of record, the largest magnitude event deposits at about 473 C.E., 605 C.E., and 1387 C.E., all occur during periods of regionally wet conditions as indicated by high lake levels. We interpret this pattern as further evidence that episodes of rapid erosion and the formation of event deposits in AL are linked to changes in the sensitivity of the landscape to erosion as influenced by changes in soil moisture and water table elevation. This interpretation is supported by regional studies of Holocene landscape evolution [*Bierman et al.*, 1997], consistent with evidence for enhanced erosion during floods in response to elevated soil moisture [*Casagli et al.*, 1999], theoretical links between climate and the occurrence of landslides [*Crozier*, 2010], and the inferred mechanism that explains the relative magnitude of erosion observed in response to TSI and the flood of 1927 [*Yellen et al.*, 2015].

The minimal response to human land cover change in comparison to the magnitude of natural variations in erosion reflects the dominant role of moisture conditions and large rainfall events in driving erosion. While human activity can make the landscape more prone to erosion, water remains the dominant trigger for landslides and means of transporting sediment. The period of most intense human impacts around AL happened to coincide with a prolonged hiatus in event occurrence, minimizing resultant erosion. In contrast, unprecedented MARs and inferred erosion rates observed in the twentieth century reflect a combination of overall wetter climatic conditions and the possible occurrence of more frequent intense rainfall events.

6. Conclusions

Sediment accumulation in AL sensitively reflects the occurrence of flooding and associated landslides and erosion in the upper Black River watershed of Vermont. While the response to the initial disturbance is rapid, elevated sediment yields may persist for decades following the initial event. Flood magnitude is not directly linked to the magnitude of erosion and subsequent deposition of event layers in AL. Significant episodes of catastrophic erosion have occurred with an average recurrence interval of ~92 years over the past 2100 years. Changes in the frequency of these catastrophic events are the primary control on long-term rates of deposition in AL and by inference, erosion of the surrounding landscape. The frequency of catastrophic erosion events recorded in AL has increased since 1900 C.E.. While human impacts cannot be completely ruled out, the frequency of destructive floods and landslides appears decoupled from human activity within the watershed and instead controlled by changes in climate and the recent occurrence of wetter conditions in the northeastern United States that favor high erosion rates and more easily trigger landslides. Thus, more frequent destructive erosion is likely under future climate scenarios calling for wetter conditions, even in the absence of an increase in flood magnitude or frequency.

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References

Appleby, P. G. (2008), Three decades of dating recent sediments by fallout radionuclides: A review, *Holocene*, *18*(1), 83–93, doi:10.1177/0959683607085598.

Barlow, M. (2011), Influence of hurricane-related activity on North American extreme precipitation, *Geophys. Res. Lett.*, *38*, L04705, doi:10.1029/2010GL046258.

Besonen, M. R., R. S. Bradley, M. Mudelsee, M. B. Abbott, and P. Francus (2008), A 1,000-year, annually-resolved record of hurricane activity from Boston, Massachusetts, *Geophys. Res. Lett.*, *35*, L14705, doi:10.1029/2008GL033950.

Bierman, P., A. Lini, P. Zehfuss, A. Church, P. T. Davis, J. Southon, and L. Baldwin (1997), Postglacial ponds and alluvial fans: Recorders of Holocene landscape history, *Geol. Soc. Am.*, *7*(10), 1–8.

Blaauw, M. (2010), Methods and code for 'classical' age-modelling of radiocarbon sequences, *Quat. Geochronol.*, *5*, 512–518, doi:10.1016/j.quageo.2010.01.002.

Brown, S., P. Bierman, A. Lini, P. T. Davis, and J. Southon (2002), Reconstructing lake and drainage basin history using terrestrial sediment layers: Analysis of cores from a post-glacial lake in New England, USA, *J. Paleolimnol.*, *28*(2), 219–236, doi:10.1023/A:1021623020656.

Casagli, N., M. Rinaldi, A. Gargini, and A. Curini (1999), Pore water pressure and streambank stability: Results from a monitoring site on the Sieve River, Italy, *Earth Surf. Processes Landforms*, *24*(12), 1095–1114, doi:10.1002/(SICI)1096-9837(199911)24:12<1095::AID-ESP37>3.0.CO;2-F.

Costa, J. E. (1975), Effects of agriculture on erosion and sedimentation in the Piedmont Province Maryland, *Geol. Soc. Am. Bull.*, *86*, 1281–1286, doi:10.1130/0016-7606(1975)86<1281:EOAOEA>2.0.CO;2.

Croudace, I. W., A. Rindby, and R. G. Rothwell (2006), ITRAX: Description and evaluation of a new multi-function X-ray core scanner, *Geol. Soc. London Spec. Publ.*, *267*, 51–63, doi:10.1144/GSL.SP.2006.267.01.04.

Crozier, M. J. (2010), Deciphering the effect of climate change on landslide activity: A review, *Geomorphology*, *124*(3), 260–267, doi:10.1016/j.geomorph.2010.04.009.

Dean, W. E., Jr. (1974), Determination of carbonate and organic matter in calcareous sediments and sedimentary rocks by loss on ignition: Comparison with other methods, *J. Sediment. Res.*, *44*(1), 242–248.

Dethier, E. N., F. J. Magilligan, C. E. Renshaw, and D. Sinclair (2015), Irene-generated hillslope failures in New England: Widespread, persistent sources of sediment to streams, *Geol Soc. Am. Abstr. Prog.*, *47*(3), 65.

Donnelly, J. P., A. D. Hawkes, P. Lane, D. MacDonald, B. N. Shuman, M. R. Toomey, P. J. van Hengstumand, and J. D. Woodruff (2015), Climate forcing of unprecedented intense-hurricane activity in the last 2000 years, *Earth's Future*, *3*(2), 49–65, doi:10.1002/2014EF000274.

Fryirs, K. A., G. J. Brierley, N. J. Preston, and M. Kasai (2007), Buffers, barriers and blankets: The (dis) connectivity of catchment-scale sediment cascades, *Catena*, *70*, 49–67, doi:10.1016/j.catena.2006.07.007.

Hoffmann, T., V. R. Thorndycraft, A. G. Brown, T. J. Coulthard, B. Damnati, V. S. Kale, H. Middelkoop, B. Notebaert, and D. E. Walling (2010), Human impact on fluvial regimes and sediment flux during the Holocene: Review and future research agenda, *Global Planet. Change*, *72*, 87–98, doi:10.1016/j.globplacha.2010.04.008.

Hooke, R. L., J. F. Martín-Duque, and J. Pedraza (2012), Land transformation by humans: A review, *Geol. Soc. Am.*, *22*(12), 4–10, doi:10.1130/GSAT151A.1.

Hubeny, J. B., J. W. King, and M. Reddin (2011), Northeast US precipitation variability and North American climate teleconnections interpreted from late Holocene varved sediments, *Proc. Natl. Acad. Sci. U.S.A.*, *108*(44), 17,895–17,900, doi:10.1073/pnas.1019301108.

Huggel, C., J. J. Clague, and O. Korup (2012), Is climate change responsible for changing landslide activity in high mountains?, *Earth Surf. Processes Landforms*, *37*(1), 77–91, doi:10.1002/esp.2223.

Kasprak, A., F. J. Magilligan, K. H. Nislow, C. E. Renshaw, N. P. Snyder, and W. B. Dade (2013), Differentiating the relative importance of land cover change and geomorphic processes on fine sediment sequestration in a logged watershed, *Geomorphology*, *185*, 67–77, doi:10.1016/j.geomorph.2012.12.005.

Marsicek, J. P., B. Shuman, S. Brewer, D. R. Foster, and W. W. Oswald (2013), Moisture and temperature changes associated with the mid-Holocene Tsuga decline in the northeastern United States, *Quat. Sci. Rev.*, *80*, 129–142, doi:10.1016/j.quascirev.2013.09.001.

Montgomery, D. R., K. M. Schmidt, H. M. Greenberg, and W. E. Dietrich (2000), Forest clearing and regional landsliding, *Geology*, *28*(4), 311–314, doi:10.1130/0091-7613(2000)28.

Noren, A. J., P. R. Bierman, E. J. Steig, A. Lini, and J. Southon (2002), Millennial-scale storminess variability in the northeastern United States during the Holocene epoch, *Nature*, *419*(6909), 821–824, doi:10.1038/nature01132.

Olson, S. A., and G. C. Bent (2013), Annual exceedance probabilities of the peak discharges of 2011 at stream gages in Vermont and selected stream gages in New Hampshire, Western Massachusetts, and Northeastern New York (No. 2013–5187), U.S. Dep. of Inter., U.S. Geol. Surv.

Parris, A. S., P. R. Bierman, A. J. Noren, M. A. Prins, and A. Lini (2010), Holocene paleostorms identified by particle size signatures in lake sediments from the northeastern United States, *J. Paleolimnol.*, *43*(1), 29–49, doi:10.1007/s10933-009-9311-1.

Pederson, N., A. R. Bell, E. R. Cook, U. Lall, N. Devineni, R. Seager, K. Eggleston, and K. P. Vranes (2013), Is an epic pluvial masking the water insecurity of the greater New York City region?, *J. Clim.*, *26*(4), 1339–1354, doi:10.1175/JCLI-D-11-00723.1.

Phillips, J. D., M. Marden, and B. Gomez (2007), Residence time of alluvium in an aggrading fluvial system, *Earth Surf. Processes Landforms*, *32*, 307–316, doi:10.1002/esp.1385.

Reimer, P. J., et al. (2013), IntCal13 and Marine13 radiocarbon age calibration curves 0–50,000 years cal BP, *Radiocarbon*, *55*(4), 1869–1887, doi:10.2458/azu_js_rc.55.16947.

Rogers, J. N. (2003), Major Holocene debris flows of Franconia Notch (White Mountains, New Hampshire) recorded in Profile Lake, MS thesis, Dep. of Geosci., Univ. of Mass. Amherst, Amherst.

Rogers, J. N., B. K. Fowler, W. D. McCoy, and P. T. Davis (2009), Post-glacial mass wasting in Franconia Notch, White Mountains, New Hampshire, in *Guidebook for Field Trips in the Northeast Kingdom of Vermont and Adjacent Regions*, edited by D. S. Westerman and A. S. Lathrop, pp. 199–223, New England Intercollegiate Geological Conference, Lyndonville, Vermont, 101st annual meeting.

Shuman, B., J. Bravo, J. Kaye, J. A. Lynch, P. Newby, and T. Webb (2001), Late Quaternary water-level variations and vegetation history at Crooked Pond, southeastern Massachusetts, *Quat. Res.*, *56*(3), 401–410, doi:10.1006/qres.2001.2273.

Trimble, S. W. (1977), The fallacy of stream equilibrium in contemporary denudation studies, *Am. J. Sci.*, *277*, 876–887, doi:10.2475/ajs.277.7.876.

Underwood, K. L., and G. Springston (2014), Alternatives Analysis Money Brook Tributary, Black River Watershed Plymouth, Windsor County, Vermont. Report prepared under contract to the Lake Rescue Association, Ludlow, VT.

Walter, R. C., and D. J. Merritts (2008), Natural streams and the legacy of water-powered mills, *Science*, *319*, 299–304, doi:10.1126/science.1151716.

Wilkinson, B. H., and B. J. McElroy (2007), The impact of humans on continental erosion and sedimentation, *Geol. Soc. Am. Bull.*, *119*(1–2), 140–156, doi:10.1130/B25899.1.

Yellen, B. C., J. D. Woodruff, T. L. Cook, and R. M. Newton (2015), Did climate change cause extreme erosion during Hurricane Irene? Expectations moving forward, *Geol. Soc. Am. Abstr. Prog.*, *47*(3), 65.

Yellen, B., J. D. Woodruff, L. N. Kratz, S. B. Mabee, J. Morrison, and A. M. Martini (2014), Source, conveyance and fate of suspended sediments following Hurricane Irene New England, USA, *Geomorphology*, *226*, 124–134, doi:10.1016/j.geomorph.2014.07.028.