HYDROLOGY, CHANNEL MORPHOLOGY, AND HOLOCENE SEDIMENTATION RECORD OF THE CENTRAL PASSAIC RIVER BASIN, NJ

by

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ABSTRACT OF THE THESIS

Hydrology, Channel Morphology, and Holocene Sedimentation Record of the Central Passaic River Basin, NJ

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The Passaic River Basin, which spans 935 mi² over northern New Jersey and parts of New York, is defined by both its glacial and post-glacial history. The retreating ice sheet (~22-18 ka) created Glacial Lake Passaic, impounded behind the Watchung Mountains until a new outlet was opened at Little Falls. Glacial rerouted the Passaic River to a northeastward course, with wetlands developing on the glacial lake sediments. One such wetland, the Great Piece Meadows, covers 2,343 acres of undeveloped floodplain within the Central Basin. Three oxbows within these wetlands, named Oxbow2, RC, and TZS, were cored to determine the flooding and geomorphological history of the central Passaic's floodplain. Radiocarbon dates and grain size data suggests that oxbow TZS was cut off from the main channel around 9-9.5 ka B.P., and that the

Oxbow2 and RC were cut off around 3.6 B.P. These dates fall within a transition from a dry to a wet period during the Holocene, which may have been a cause for the avulsions. Using mercury soil concentrations, deposition from the past ~180 years were determined to be within the upper 5-17 cm of the cores. The Passaic's central and lower basin have long been plagued by flooding problems, with the worst flood on record occurring in 1903. Analysis of peak discharge at gages along the Passaic and its tributaries for 27 major floods confirms observed flashy discharge of the tributaries and the backup of floodwaters from Little Falls to the central basin during flooding. While the Passaic Basin's morphology makes it prone to flooding, there are anthropogenic factors as well. While peak discharges for major floods in the Passaic have decreased since the turn of the century, major flood frequency is increasing. Reservoirs are shown to significantly decrease annual runoff and flood discharge along rivers directly downstream from them, but while they can reduce the intensity of flooding, they do not prevent major floods. Annual average runoff ratio values have increase for most areas of the Passaic Basin's streams as well, despite the construction of reservoirs and diversions for water supply. This implies that an increase in urban and suburban development is a factor in the river's increasing flooding problems.

ACKNOWLEDGEMENTS

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INTRODUCTION

The Passaic River, located in northeast New Jersey has a watershed of about 935 square miles (Hofman, 1955) (Fig. 1), and runs through Passaic, Bergen, Essex, Morris, Union, and Somerset, and Hudson Counties before emptying out into Newark Bay. The basin also includes parts of Sussex County in New Jersey, and Rockland and Orange Counties in New York. Its four tributaries are the Rockaway and Whippany Rivers, which join at Pine Brook, the Pompton River, which joins below Two Bridges, and the Saddle River, which joins downstream of Paterson (Hofman, 1955; Hollister and Leighton, 1903). One of its main tributaries, the Pompton River, is fed by the Wanaque, Ramapo, and Pequannock Rivers located in the Highlands. All tributaries except the Saddle River connect to the Passaic within the Central Basin before the river flows through the gap in the Second Watchung Mountain at Little Falls (Hollister and Leighton, 1903). The Passaic River Basin has a complex history spanning from ~150,000 years ago to the past century. Blockage of its original course by ice and moraine deposits gave the Passaic River a long, low gradient course that flows to the northeast before turning southward around Paterson. .(Stanford, 2007a). Glaciation also cased the formation of Glacial Lake Passaic, which left buried valleys within the bedrock filled with lake deposits. The extensive wetlands of the Central Basin have developed on these lake deposits (Hoffman, 1989).

Chronic flooding problems have characterized the Passaic River Basin's post-colonial history. Flooding issues have been recorded in the Passaic Basin since colonial times, especially along the Central and Lower Passaic (Brydon, 1974). The Passaic River's morphology has made it naturally prone to flooding. During storms, water rushing down from the steep gradient tributaries hits the gently sloping course of the Passaic within the Central Basin, and floodwaters back up from the narrow gap in the basalt bedrock at Little Falls and into the wetlands of the Great Piece Meadows in Fairfield. Glacial lake deposits underlying the Great Piece Meadows further results in poor drainage and floodplains with widespread wetlands. This causes major flooding both up and downstream, with the largest flood to on record being the flood of 1903, a 100-year flood event.

Though the river is naturally prone to flooding, human impacts on the basin may be a factor in the river's increasing destructive floods. Industrial cities along the Passaic and the proximity to New York have caused vast expansions of population and urbanization since Europeans settlement. This surge in urbanization over the past two centuries has resulted in a loss of wetlands and forested areas and an increase in impermeable surfaces, while increases in demand in water supply has led to river diversions and reservoir construction. It is important to examine both the pre-settlement history of the Central Passaic floodplain during the Holocene and the post-settlement history of the Passaic Basin to better understand the factors that shape the behavior of the Passaic River. One large area of wetlands, the Great Piece Meadows, along the central Passaic River at Fairfield, contains several oxbow cutoffs. Because of their retention of flood deposits and their unlikely existence in low gradient floodplain, these oxbows offer insight into the flooding history and effects of flooding along the central Passaic during the Holocene. Flood, water supply, precipitation, and stream gage records from the past ~200 years provide recent information about the nature of flooding and human impacts on the basin. Examining flooding patterns, flood frequencies, and runoff trends of the

Passaic and its tributaries gives insight into the factors that cause, exacerbate, and influence contemporary flooding, while floodplain stratigraphy provides information about Holocene flooding and river channel changes in the central Passaic River.

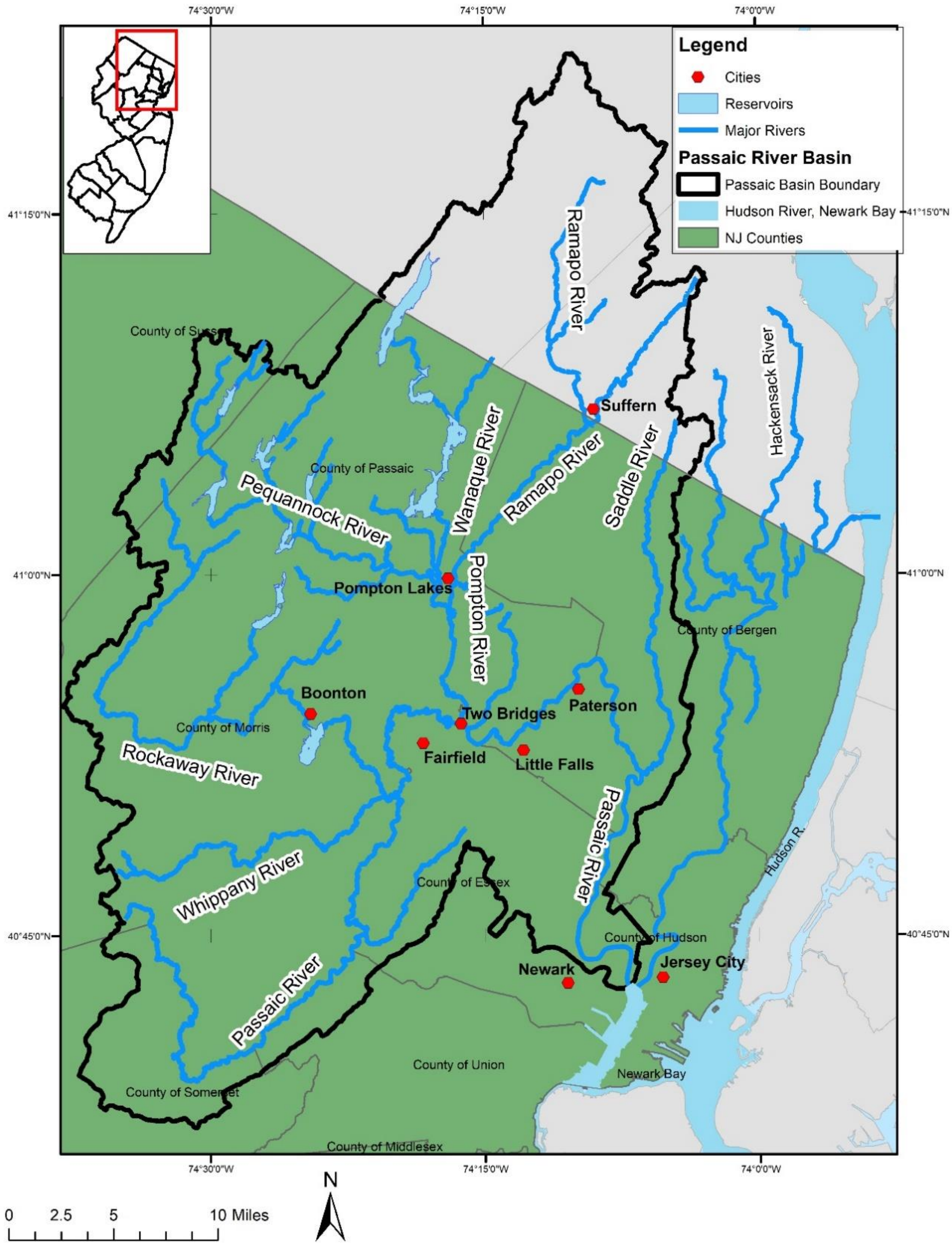


Fig 1: Passaic River Basin. Data from (NJDEP et al., 2010b; NJOIT and NJOGIS, 2017)

OBJECTIVES

The three main objectives of this project are: 1- To establish the Holocene flooding history and sedimentation record of the Central Passaic River Basin by analyzing sediment deposits of cores taken from three oxbows. 2- To characterize the behavior of the Passaic River and its tributaries during major flooding within the Passaic River Basin over the past 200 years using stream gage data and past reports. 3- To discuss changes in flooding patterns and runoff since the turn of the century, and possible factors that caused changes.

HYPOTHESIS

While the Passaic River is naturally prone to flooding within the Central and Lower Basins, a variety of factors influence the magnitude and frequency of flooding. Floods have been a problem within the Passaic Basin since colonial times, but flooding has appeared to become more problematic and destructive since the turn of the century. Because of the power provided by the river and the proximity to New York, the Passaic River Basin has been heavily developed. The process of urbanization is known to increase flooding issues. I hypothesize that flooding patterns and deposits of the central Passaic River after European settlement vary from those of the Holocene, which can be determined from floodplain sediment, and recent discharge and flood records of the Passaic and its tributaries. Through analysis of discharge of the Passaic and its tributaries as well as climate data, land use, and water use data, I plan to test the hypothesis that urbanization has had an impact on flooding within the Passaic River Basin, and is a factor in recent flooding problems.

GEOLOGY

BEDROCK GEOLOGY

The bedrock of the Central Passaic Basin is made up of four different formations. The Preakness basalt, which underlies the Second Watchung Mountain, has an estimated thickness of ~ 984 feet (300 meters) (Drake et al., 1996; Hoffman, 1989)(Fig. 2). The Passaic River is joined by the Pompton River before exiting through a gorge in Second Watchung Mountain at Little Falls and continuing through another narrow gorge in First Watchung Mountain at Paterson. During high flow periods, these narrow gorges impound the river and lead to flooding upstream; some of the most severe flooding in the Passaic River Watershed (Hollister and Leighton, 1903). The lower 70 feet of this formation has numerous vesicles (Drake et al., 1996). The Hook Mountain basalt, which underlies Hook Mountain and Long Hill, has a maximum thickness of about 360 feet (109.7 meters). The Towaco Formation, which underlies the lowland between the Second Watchung Mountain and Hook Mountain, includes mudstone, shale, sandstone, and siltstone, and is roughly 1115 feet (340 meters) thick (Hoffman, 1989). The bedrock aquifer, made up of fractured sedimentary rocks and interbedded basalt units, is significantly less productive than the surficial aquifer (Hoffman, 1989). The Boonton Formation is made up of sandstone, siltstone, and silty mudstone and underlies the lowland west of Hook Mountain and Long Hill (Drake et al., 1996; Owens et al., 1998). The Lower Jurassic Feltville Formation is comprised of interbedded sandstone with siltstone and mudstone. Separated by the Ramapo Fault, the bedrock of the Highland region is significantly different from the sedimentary bedrock of the Central Basin and Lower Valley (Fig.2).

The bedrock of the Highlands is composed of metamorphic rocks, such as gneiss and marble, as well as igneous rock (mainly granite) (Drake et al., 1996). The igneous and metamorphic bedrock of the highlands is difficult to erode, causing its rivers to be narrow with high gradients (Hollister and Leighton, 1903).

A.

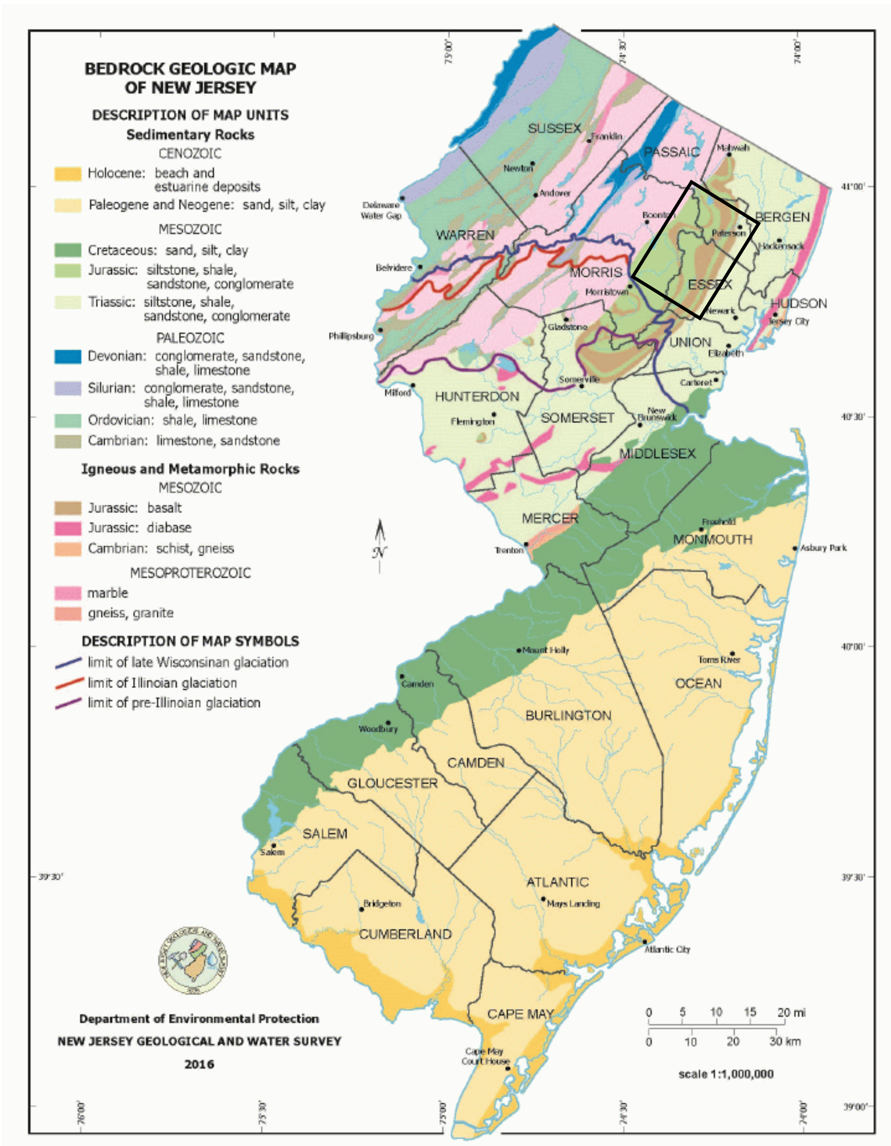
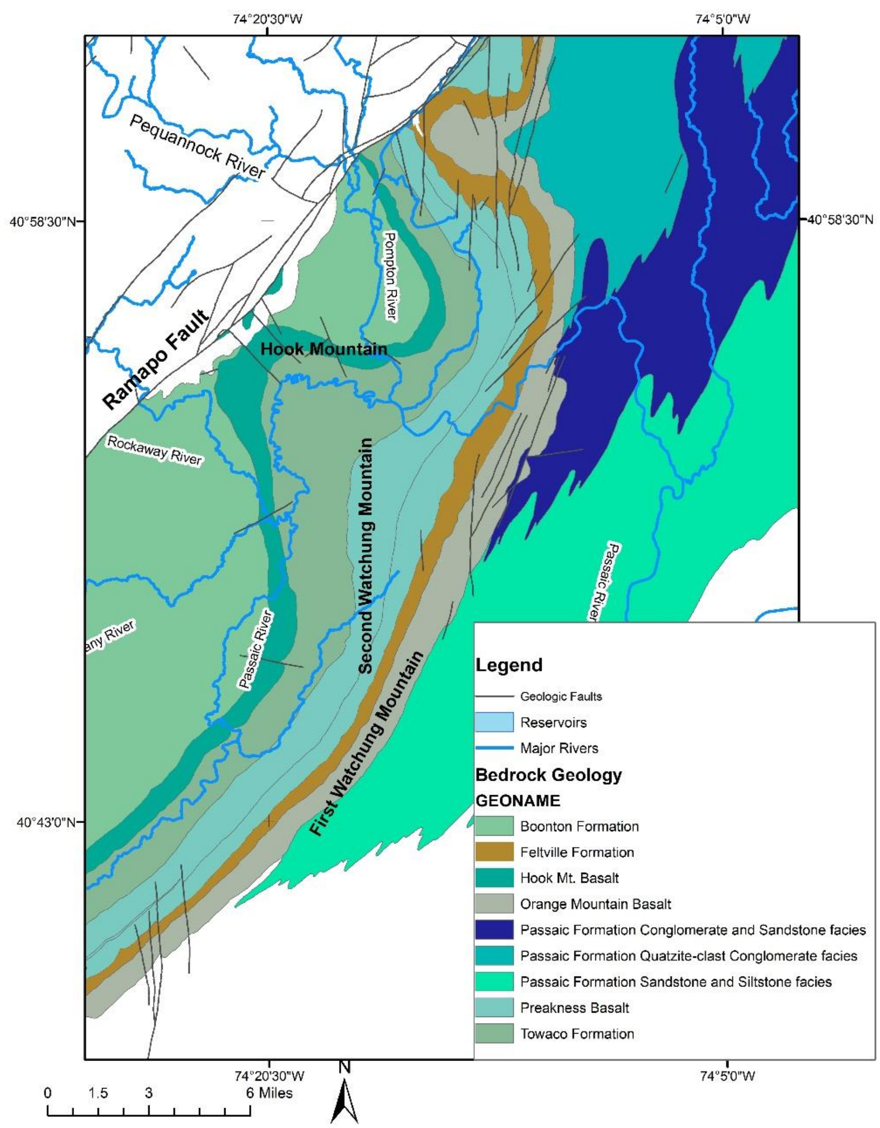


Fig. 2A: Bedrock geology of New Jersey and an enlarged map of the area within the black box showing bedrock geology of the Central Basin and Lower Valley (right) (Dalton et al., 2014; Drake et al., 1996)

B.

Fig. 2B: An enlarged map of the area within the black box showing bedrock geology of the Central Basin and Lower Valley. Data from (Dalton et al., 2014; Monteverde et al., 2015; NJDEP et al., 2010b)



SURFICIAL GEOLOGY

The Passaic River Basin contains a network of interconnected depressions within the bedrock known as buried valleys (Hoffman, 1989). These depressions were scoured out by glaciers, and are thought to be the course of the pre-glacial Passaic River. This study focuses on cores taken from the alluvial deposits overlying the Fairfield buried valley, which covers the southern Pompton Plains Quadrangle and the northern Caldwell Quadrangle. The aquifer system of the Central Passaic River Basin is split into a surficial and a bedrock aquifer, although there may be several waterbearing zones existing within each aquifer (Hoffman, 1989). The more productive surficial aquifer consists of the alluvial and glacial sediments which fill the area's interconnected buried valleys (Hoffman, 1989)

The general top surficial layer of this area is made up of alluvium deposited along the floodplain, wetlands, and channels of the Passaic and Pompton rivers (Stanford, 2007b). This alluvium is made up of fine to medium sand, pebbly sand, silty sand, silt, and clayey silt, as well as organic matter (Fig. 3). The sand and silty sand are mainly found in natural channel levees and as overbank deposits of the river, while the silt and clay sediments are found farther away from the rivers in the backswamp areas (Stanford, 2007b). These deposits can be up to 20 ft (6.1) meters thick, but thin out towards Troy Meadows and Hatfield Swamp in the west. Further away from the river channels, stream terrace deposits of fine-medium sand, and pebbly sand to pebbly gravel, lie on top of the glacial lake deposits and can be up to 30 feet thick, but usually range from five to 15 feet (1.5-4.57 meters) (Stanford, 2007b). These deposits make up the upper unconfined unit of the aquifer system.

A.

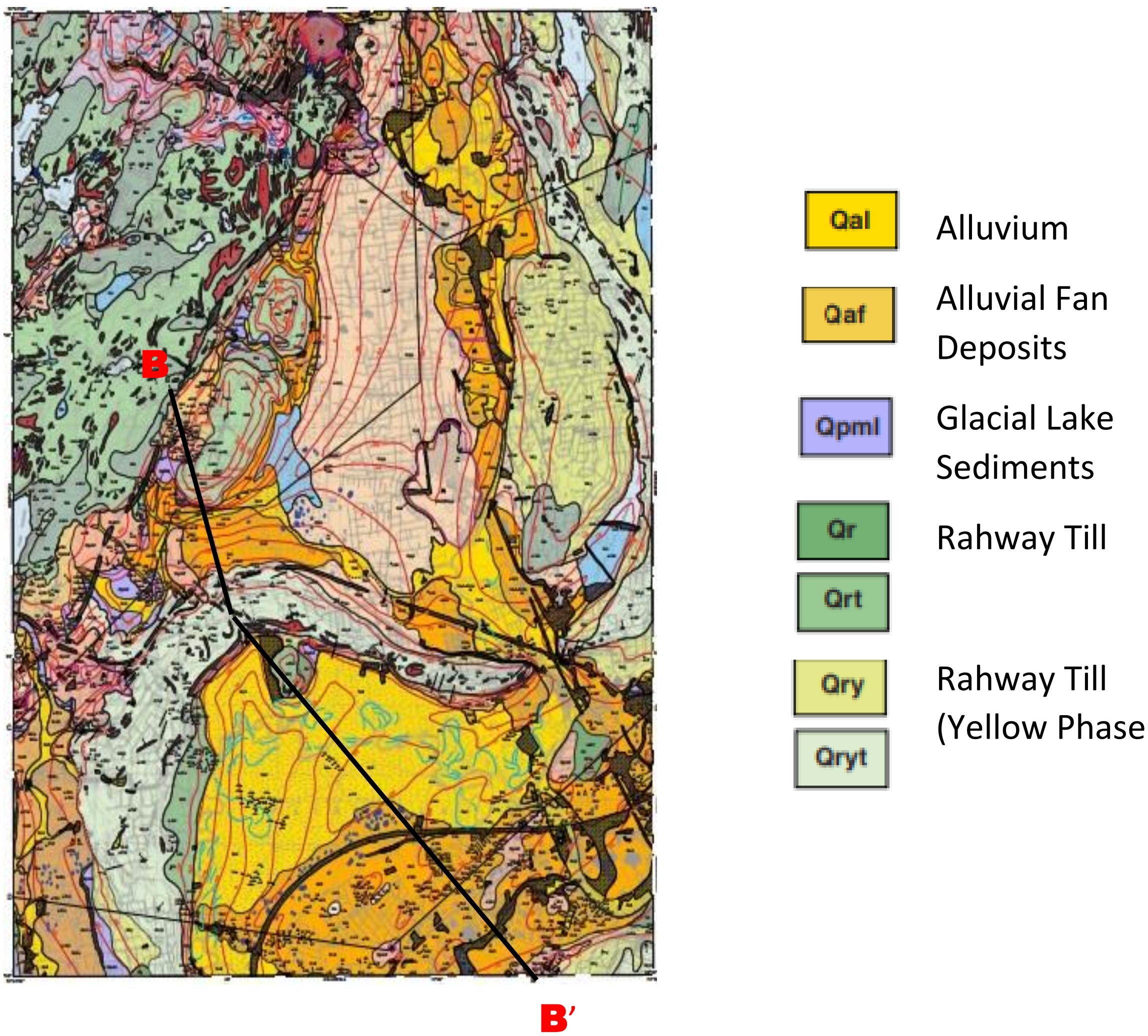
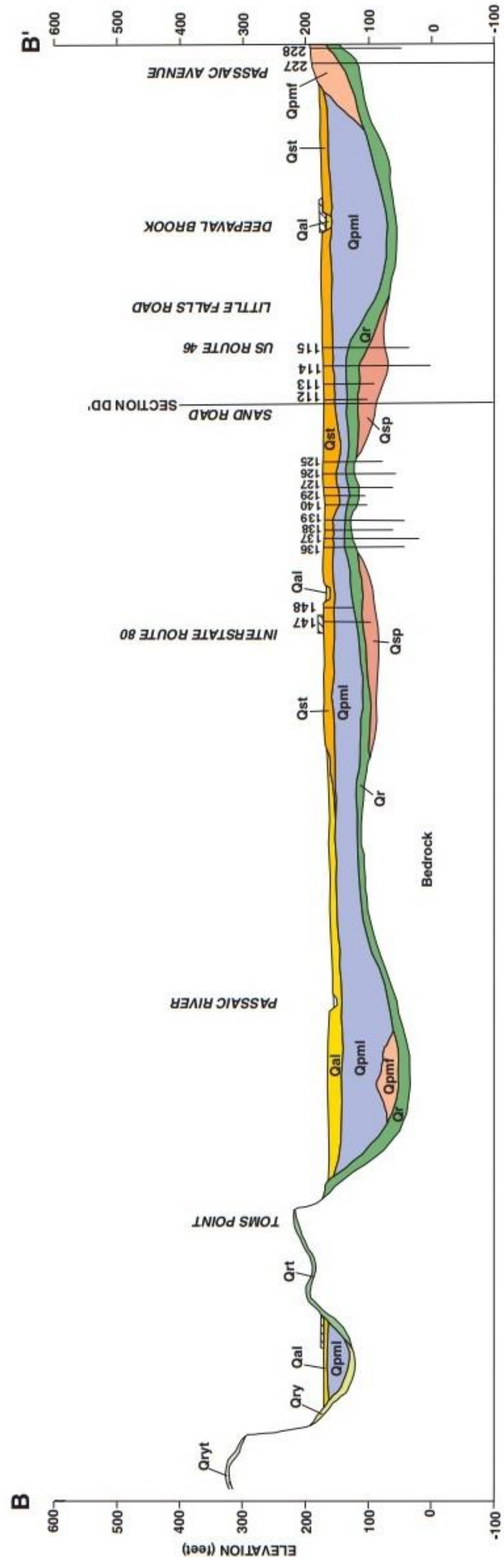


Fig. 3: Surficial geology of the Pompton Plain Quadrangle within the Central Passaic River Basin and abbreviated cross section of the bend in cross section B to B′ , with 10x vertical exaggeration. The Pompton Plains Quadrangle includes the location of the study area (Stanford, 2007b)

B.


B
B'
ELEVATION (feet)
600
500
400
300
200
100
0
-100
PASSAIC AVENUE
DEEPAVAL BROOK
LITTLE FALLS ROAD
US ROUTE 46
SAND ROAD
SECTION DD'
INTERSTATE ROUTE 80
PASSAIC RIVER
TOMS POINT
Bedrock
228
227
115
114
113
112
125
126
127
129
140
139
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137
136
148
147
Qpmf
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Qal
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Qryt

Glacial till in this area poorly sorted and generally compact below the soil zone. The main glacial till deposits in this area are Rahway till, the Bergen till, and the Netcong till (Stanford, 2007b). The Rahway and Netcong tills are of the same age; both being deposited in the Late Wisconsinan (150 ka). The Bergen till is older, probably dating from the Illinoian. The Rahway Till is reddish brown to reddish yellow silty sand to sandy silt although it also contains pebbles, cobbles, and a few boulders of gneiss, basalt, sandstone and quartzite (Stanford, 2007b). The Rahway Till, found overlaying Hook Mountain, is usually about 9-3 meters thick, but can be as thick as 45 meters on the south east end of the mountain. The Netcong Till is a yellow-brown sandy-silt to silty-sand with gneiss, quartzite, and conglomerate boulders and cobbles. Although it can be as much as 30 meters thick, it is usually only about six meters (Stanford, 2007b). This glacial till is found in thin deposits on top of the Watchung Mountains.

Glacial lake bottom deposits consist of very fine to fine sand, silt, and clay deposited mainly in the Moggy Hollow stage of Glacial Lake Passaic (Stanford, 2007b). The lake deposits are some of the thickest surficial deposits in the area, reaching up to 260 feet (79 meters) thick. These fine grained lake bottom sediments form a confining unit which greatly hinders the vertical flow of water from the wetlands to the deeper aquifers. Because of this, the wetlands of the Passaic usually do not affect the lower aquifers of the groundwater system (Hoffman, 1989)

GLACIAL HISTORY

GLACIATION AND GLACIAL LAKES

The current course of the Passaic River was mainly influenced by the Illinoian and Wisconsinan glaciations. Before the Illinoian glaciation, which occurred about 150,000 years ago, the modern Passaic River was drained by two separate river systems.

The ancestral Passaic River flowed out from the gaps in the Watchung Mountains at Millburn and Short Hills and fed the ancestral Raritan River. It is likely that the Pompton River also exited the gap at Short Hills as a tributary to the Passaic (Stanford, 2007b). Sediments deposited by the glacier partially filled the gap at Short Hills and the surrounding valleys, which did not alter the river's course at the time but contributed to the outlet's eventual obstruction (Fig. 4A).

About 125,000 years after the Illinoian glacier retreated, the Late Wisconsinan Glaciation extended ice into the Passaic Valley. During the Late Wisconsinan Glaciation, which occurred about 22,000-18,000 years B.P., ice advanced into the Passaic Valley east of the Watchungs (the Hackensack Lobe) and west of the Watchungs (the Passaic Lobe) (Stanford, 2007a). Because of the mountainous terrain in the Highlands, the glacier advanced more slowly in the Highlands than in the Central Basin and Lower Valley (Stanford, 2007b). Ice from the Hackensack Lobe blocked the outlet at Millburn Gap, cutting off drainage from the Passaic Valley. Inflowing water, blocked from draining and bounded by the glacier to the north, the escarpment of the Ramapo fault to the west, and the Watchungs to the south and east, created Glacial Lake Passaic (Cavallo et al., 1994; Salisbury and Kümmel, 1895) (Fig. 4B). While Glacial Lake Passaic existed for about

1,500 years, the advance and recession of the glacier caused alterations, leading to different stages of the lake (Stanford, 2007a) (Fig. 5).

A.

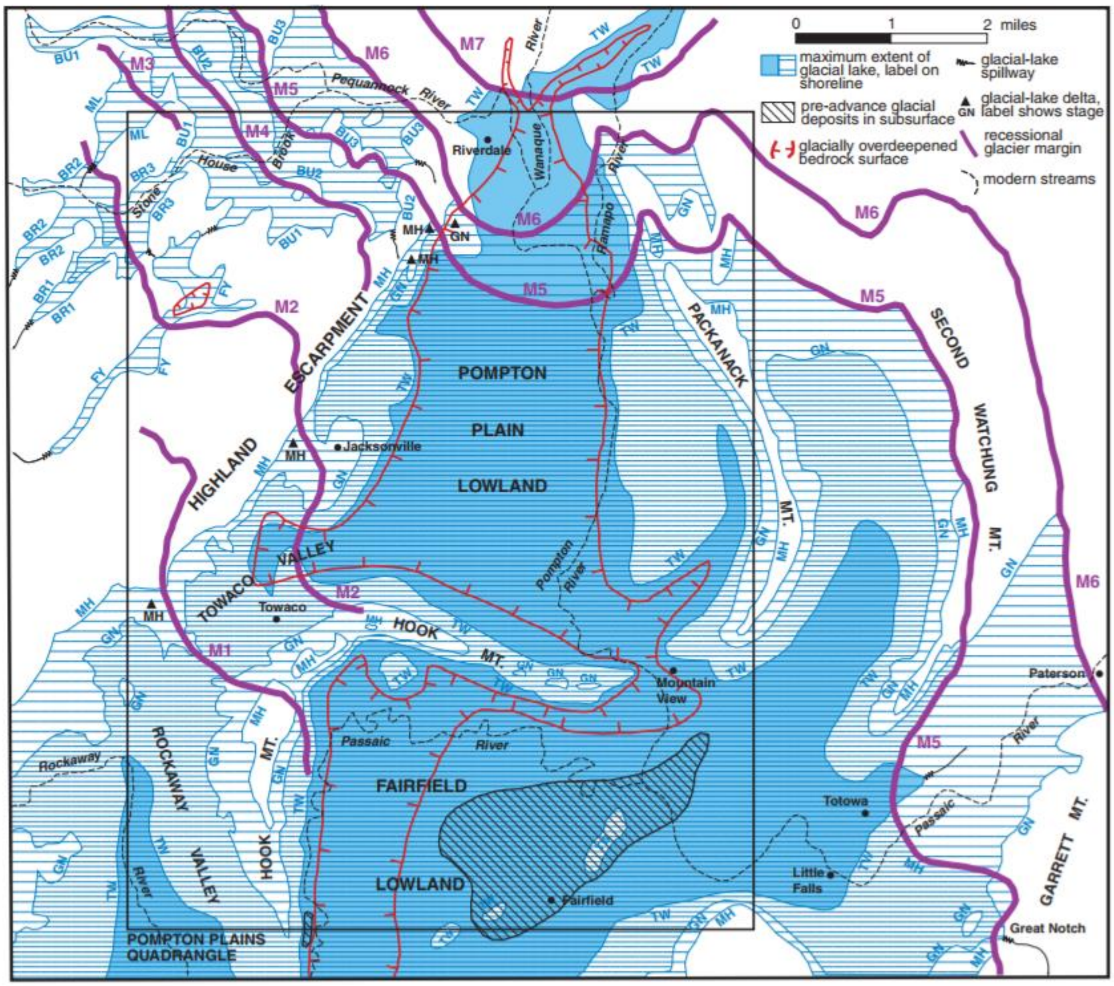
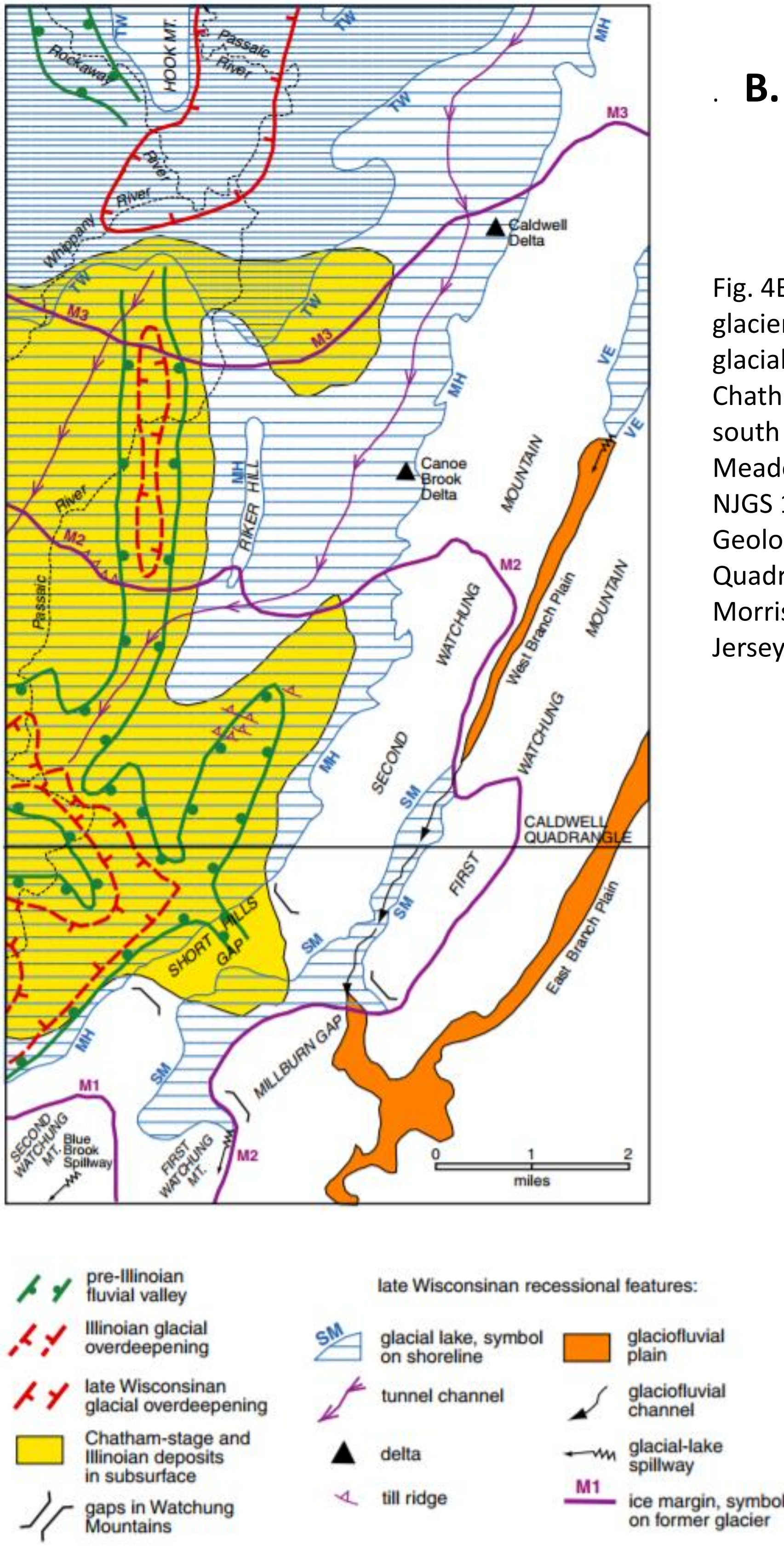


Fig. 4A: Map of the glacier margins and the extent of glacial lake shorelines within the Pompton Plains Quadrangle in the Central Passaic River Basin. (NJGS 1:24000 Surficial Geology of the Pompton Plains Quadrangle Morris, Passaic, Essex, and Bergen Counties, New Jersey (Stanford, 2007b)



B.

Fig. 4B: Geomorphic, glacier deposits, and glacial lake deposits of the Chatham Quadrangle, south of the Great Piece Meadows, taken from NJGS 1:24000 Surficial Geology of the Caldwell Quadrangle Essex, and Morris Counties, New Jersey (Stanford, 2005)

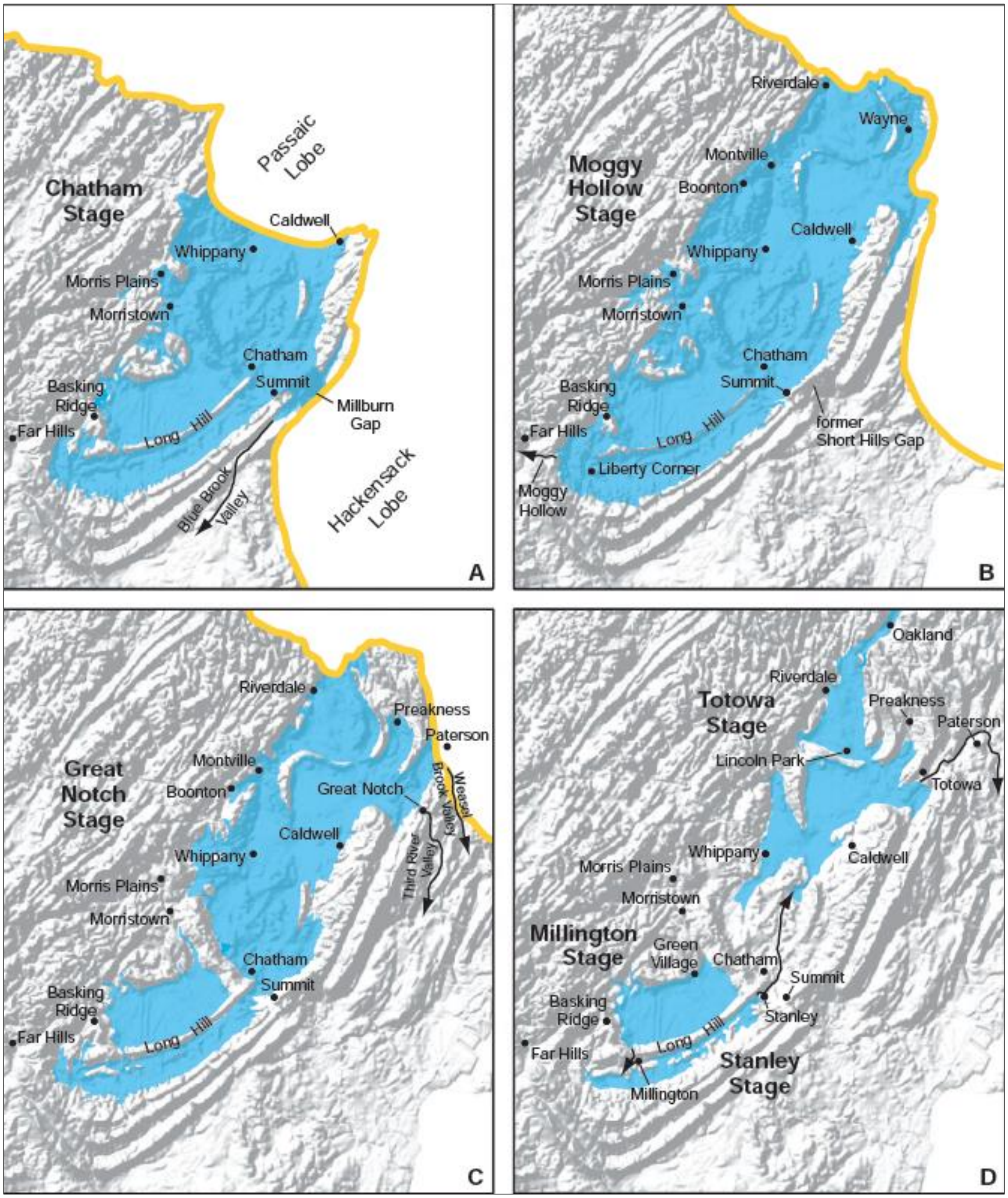


Fig. 5: The Chatham, Moggy Hollow, Great Notch, and the Totowa, Millington, and Stanley Stage of Glacial Lake Passaic (Stanford, 2007a)

The first stage of the lake was the Chatham stage, in which the lake was up to 200 feet deep (Fig. 5A). Elevation along the shoreline during this stage ranged from 250-300 feet. This stage of the lake drained through the Blue Brook Valley (Stanford, 2007a). As the glacier advanced, drainage through the Blue Brook Valley was blocked, causing the lake to rise to the Moggy Hollow stage (Fig.5B). The Moggy Hollow stage was controlled by a spillway across a low point on the crest of Second Watchung Mountain, creating a mile long sluiceway, now known as Moggy Hollow. During the retreat of the glacier, terminal moraine deposits completely blocked the gap at Short Hills allowing Glacial Lake Passaic to extend northward to where Wayne is currently located and forming a lake up to 500 feet (Salisbury and Kümmel, 1895). Meltwater streams from the receding glacier formed deltas along the shore of the lake Shoreline elevations were around 340 feet at the southern end, and 410 feet at its northeastern end.

Retreating ice opened up a gap in the First Watchung Mountain in Clifton, NJ, causing the lake to lower 80 feet to the Great Notch stage (Fig. 5C) (Salisbury and Kümmel, 1895; Stone et al., 2002). The Great Notch stage had a maximum depth of 300 feet. Shoreline elevations were at 260 feet to the south, and 320 feet in the north. Receding ice eventually opened another outlet in the First Watchung Mountain near present day Paterson, resulting in the sudden drainage of the Great Notch stage. Rapid flooding from this drainage cut a sluiceway that currently comprises part of the Weasel Brook Valley in Paterson.

POST GLACIAL LAKES

During the glacial period of the Passaic River basin, the earth's crust in that area was depressed from the weight of the ice, especially in in the north (Cavallo et al., 1994; Stone et al., 1989). After the retreat of the glacier, the area rebounded, but differential isostatic rebound caused the basin to ultimately rise to the north at a gradient of 0.0004 toward N28°E (Stone et al., 1989; Stone et al., 2002). Due to the combination of the draining of Glacial Lake Passaic and rebound, widespread erosion and stream incision of glacial deposits within the basin occurred (Cavallo et al., 1994). After the drainage of the Great Notch stage of Glacial Lake Passaic, several postglacial lakes were formed due to damming by moraine and lacustrine fan deposits (Stanford, 2007a). What was formerly Glacial Lake Passaic was divided into smaller lakes: the Totowa stage, the Stanley stage, and the Millington stage (Fig. 5D). The largest stage of these lakes, the Totowa stage, formed when glacial sand and gravel deposits blocked the valley at Totowa. The shoreline elevation varied along the lake, with its southern shore at Whippany at 170 feet, its northern shore at Oakland at 220 feet, and it's eastern shore near Totowa at 190 feet

What is currently Pompton Plains lies atop a large delta formed in the Totowa stage of the lake. Sand terraces that flank the Passaic, Whippany, and Rockaway Rivers in the Central Basin were also deposited into the Totowa stage lake (Stanford, 2007b). The Stanley stage, which is south of the Totowa, was formed by terminal moraine deposits damming Passaic Valley between Chatham and Summit. This stage was much smaller than the Totowa, and only had a depth of 30 feet. Both the northern and southern shoreline elevations were similar, ranging from 220-230 feet. The Millington stage occupied the area of what is now the Great Swamp, which, prior to the late Wisconsinan glaciation, had been a valley draining to the Short Hills Gap (Stanford, 2007a). This drainage outlet was filled in by delta and moraine deposits from the Moggy Hollow stage. The Millington stage is separated from the Stanley stage by Long Hill, and has a maximum depth of 50 feet. Its southern shore at Basking Ridge had an elevation of 260 feet, similar to its northern shore's elevation of 270 feet.

Both the Totowa and the Stanley stage eventually broke through the moraine deposits blocking their outlets and completely drained (Stanford, 2007a). Drainage from the Millington stage, however, eroded through the basalt bedrock and carved out a gorge in Long Hill until the lake completely drained about 14,000 years ago. Millington Gorge, as it is now called, became part of the route of the postglacial Passaic River (Salisbury and Kümmel, 1895). Because of the damming of the Short Hills Gap, the filling in of the ancestral Passaic River Valley by glacial deposits and glacial lake sediment, and postglacial rebound that elevated the northern parts of the central basin relative to southern parts, the postglacial Passaic River developed a low gradient, northeasterly route in order to exit the basin through the gaps at Little Falls and Paterson (Salisbury and Kümmel, 1895; Stanford, 2007a). The Little Falls gap, which was revealed after the retreat of the ice, is at a higher elevation than the former outlet at Short Hills, and resulted in the addition of about 740 sq. miles of drainage area to the Lower Passaic River (Salisbury and Kümmel, 1895). The river's meandering path also features straight stretches and angular meander bends, which is indicative of rivers underlain by fine grained sediments (Cavallo et al., 1994; Hofman, 1955).

GEOMORPHOLOGY

THE PASSAIC RIVER AND TRIBUTARIES

The Passaic River Drainage Basin, which covers about 935 mi^2 of Northern NJ and Southern NY (Hofman, 1955) is divided into three main hydrologic regions; the Highlands, the Central Basin, and the Lower Valley (Fig. 6).

The Highlands region is located in the mountainous northern reaches of the drainage basin and stretches about 15 miles wide (Hofman, 1955). The maximum elevation in this area is 1,300 feet above sea level. The major rivers of the Highlands flow through steep sided valleys in the primarily metamorphic bedrock (Hollister and Leighton, 1903). Because of the high gradient of the rivers and low permeability of the metamorphic bedrock, runoff from the Highland rivers is usually flashy (Fig. 7).

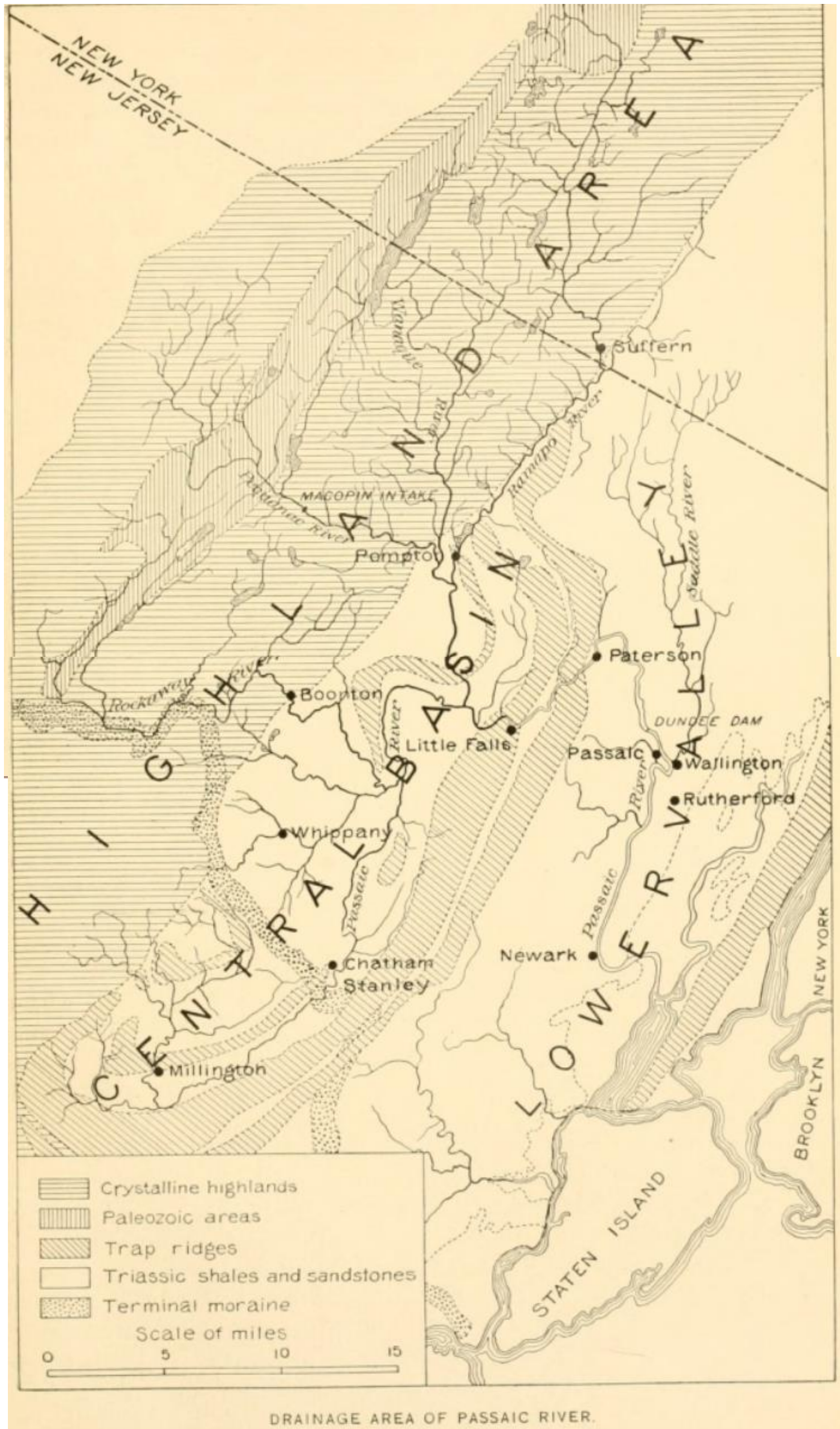


Fig. 6: Regions of the Passaic River Basin (Hollister and Leighton, 1903)

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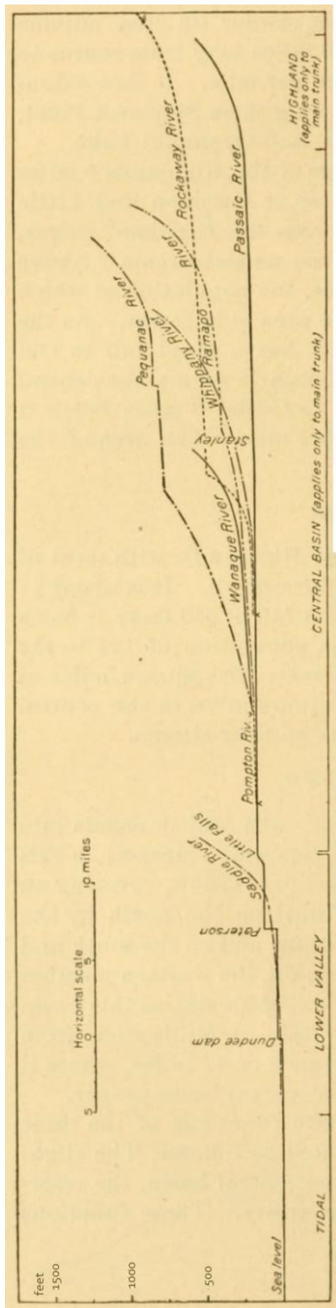

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Fig. 7: A general cross section of the Passaic River Basin, showing the relative gradients of the Passaic River and its tributaries. (modified from Hollister and Leighton, 1903)

The Central Basin stretches about 30 miles from the Great Swamp in the south to Pompton Lakes in the north, following the Passaic River from Morris County to the outlet in the Watchung Mountains at Little Falls (Hofman, 1955). The low point of the Central Basin lies about 180 feet asl, and is underlain by sedimentary bedrock which is more easily eroded than the igneous and metamorphic rocks of the highlands (Hollister and Leighton, 1903). The Passaic River has a low gradient, slow moving flow until it reaches the 40 foot drop through the Second Watchung Mountain at Little Falls. Multiple expanses of wetlands also characterize the Central Basin, that typically store floodwaters after flooding events (Brydon, 1974; Hollister and Leighton, 1903). The Lower Valley extends from the outlet at Little Falls to the mouth of the river at Newark Bay (Hofman, 1955). After the drop at Little Falls, the river follows a gently sloping, meandering path until a 70 foot drop in the First Watchung Mountain at the Great Falls in Paterson (Hollister and Leighton, 1903). The river then continues to Dundee Dam, from which it proceeds to drop six feet in elevation over the next four miles. The last 13.5 miles to the Newark Bay is a tidal estuary. Floods in the Lower Passaic are also affected by high tides, which further combines with peak river discharges to produce some of the worst flooding along the Passaic (Brydon, 1974).

The Passaic River has four main tributaries; the Pompton, the Rockaway, the Whippany, and the Saddle Rivers (Hollister and Leighton, 1903)(Fig.1). The Whippany River is a fast moving stream about 17 miles long, and mainly passes through the glacial moraine sediments of the hills north of Morristown. Three tributaries, the Whippany, the Rockaway, and the Pompton, feed the Upper and Central Passaic River. While they originate in the Highland region the Whippany and the Rockaway join the Passaic River at Pine Brook, just before the path of the river bends at Hook Mountain within the Great Piece Meadows. The Pompton joins the Passaic further downstream at Two Bridges. The Pompton's tributaries, the Wanaque, Pequannock, and Ramapo Rivers, are influenced by precipitation in the Highlands region of the basin, and contribute to the total flow of water into the Passaic River at Two Bridges (Fig. 8).

The Ramapo River originates in New York, and enters New Jersey at Suffern (Hollister and Leighton, 1903). The Ramapo has a steep gradient of 3.6 x 10^{-3} as it cuts though the crystalline bedrock in New York but flattens to 1.1 x 10^{-3} ft/mile when it reaches the sandstone bedrock in Northern New Jersey. At Pompton, it passes over a dammed natural waterfall before joining the Wanaque and Pequannock Rivers to form the Pompton River (Hollister and Leighton, 1903). Saddle River joins the Passaic downstream from Paterson, but is not taken into account in influencing flooding of the Central Passaic River because it is located in the Lower Valley, downstream of the drop at Little Falls. Discharge data from 11 stream gages located along the Passaic, its tributaries, and the Pompton River's tributaries were used to analyze streamflow patterns during major floods (Table 1).

The Great Piece Meadows are especially known to frequently flood. Discharge from the narrower tributaries in the Highlands flows into the wide floodplain of the Central Passaic (Hollister and Leighton, 1903). While the surficial alluvium along the floodplain is permeable, the thick glacial lake clay layer, about 7-15 feet below the surface has a low vertical and horizontal conductivity (Hoffman et al., 1989; Stanford, 2007b). Incoming floodwaters from the adjoining tributaries cannot be fully absorbed by the floodplain sediments, and pools up within the wetlands of the Great Piece Meadows

(Brydon, 1974). The wetlands, however, soak up and retain some of the floodwaters, acting as a natural buffer to flooding directly downstream and in the Lower Passaic River ((Hollister and Leighton, 1903; US Army Corps of Engineers, 1987). Close downstream of the Great Piece Meadows, the Passaic passes over Preakness basalt at Little Falls. This outlet is narrow, and the basalt rock makes is difficult to erode and widen the channel.

RIVER GRADIENTS

The difference of the Passaic River's gradient compared to those of its tributaries is a significant factor in the Passaic's major floods. The Whippany, Rockaway, Ramapo, Wanaque, and Pequannock Rivers originate in and flow primarily through the mountains of the Highlands, while the main course of the Passaic transverses the much flatter, lower elevation area of the Central Basin (Hollister and Leighton, 1903). The Highland tributaries are also narrower because their underlying metamorphic and igneous bedrock is harder to erode than the sedimentary bedrock of the Central Basin. The Pompton River begins at the boundary of the Highlands region and Central Basin, giving it a much shallower gradient than its tributaries. Hollister and Leighton's report on the flood of 1902 lists the Passaic River as having an average gradient of 1.3 x 10^{-3}, while the Pompton River has an average gradient of only 3.1 x 10^{-4} (Table 1). The tributaries of the Pompton and Passaic are much steeper in comparison, with average gradients ranging from 4.7 x 10^{-3}- 6.6 x 10^{-3} ft/mi for the Passaic's tributaries, and a range of 3.4 x 10^{-3} - 6.6 x 10^{-3} ft/mi for the Pompton's tributaries (Table 1) (Hollister and Leighton, 1903). This sharp contrast in gradients has been known to make the Central Basin prone to flooding (Hollister and Leighton, 1903). When heavy precipitation in the Highlands washed in by these steep, narrow tributaries suddenly hit the almost flat gradients of the Passaic River at Pine Brook and the Pompton at Pompton Plains, the low gradient channels are unable to contain the flow, and flood waters spill out along and beyond the floodplain. The Pompton further carries the flow of its Highland tributaries to Two Bridges, where it joins with the Passaic, adding greatly to the Passaic's peak discharges during flooding (US Army Corps of Engineers, 1987). The large influx of water is unable to drain easily because of the constriction of the narrow outlet in the basalt rock at Little Falls, further contributing to widespread flooding in both the Central and Lower Passaic River.

River	Drainage Area (mi²)	Length of River (mi)	Avg. Gradient (ft/mi)	Avg. Gradient
Passaic River	299	85.5	6.9	0.0013
Saddle River	60.7	17	31.8	0.006
Pompton River	24.8	6	1.66	0.00031
Ramapo River	160.7	34	18.8	0.0036
Wanaque River	109.6	16.5	33.9	0.0064
Pequannock River	84.8	31	35.1	0.0066
Rockaway River	138.4	40	25.7	0.0049
Whippany River	71.1	20	34	0.0064

Table 1: Drainage areas and average gradients of the Passaic and its tributaries. Revised from (Hollister and Leighton, 1903)

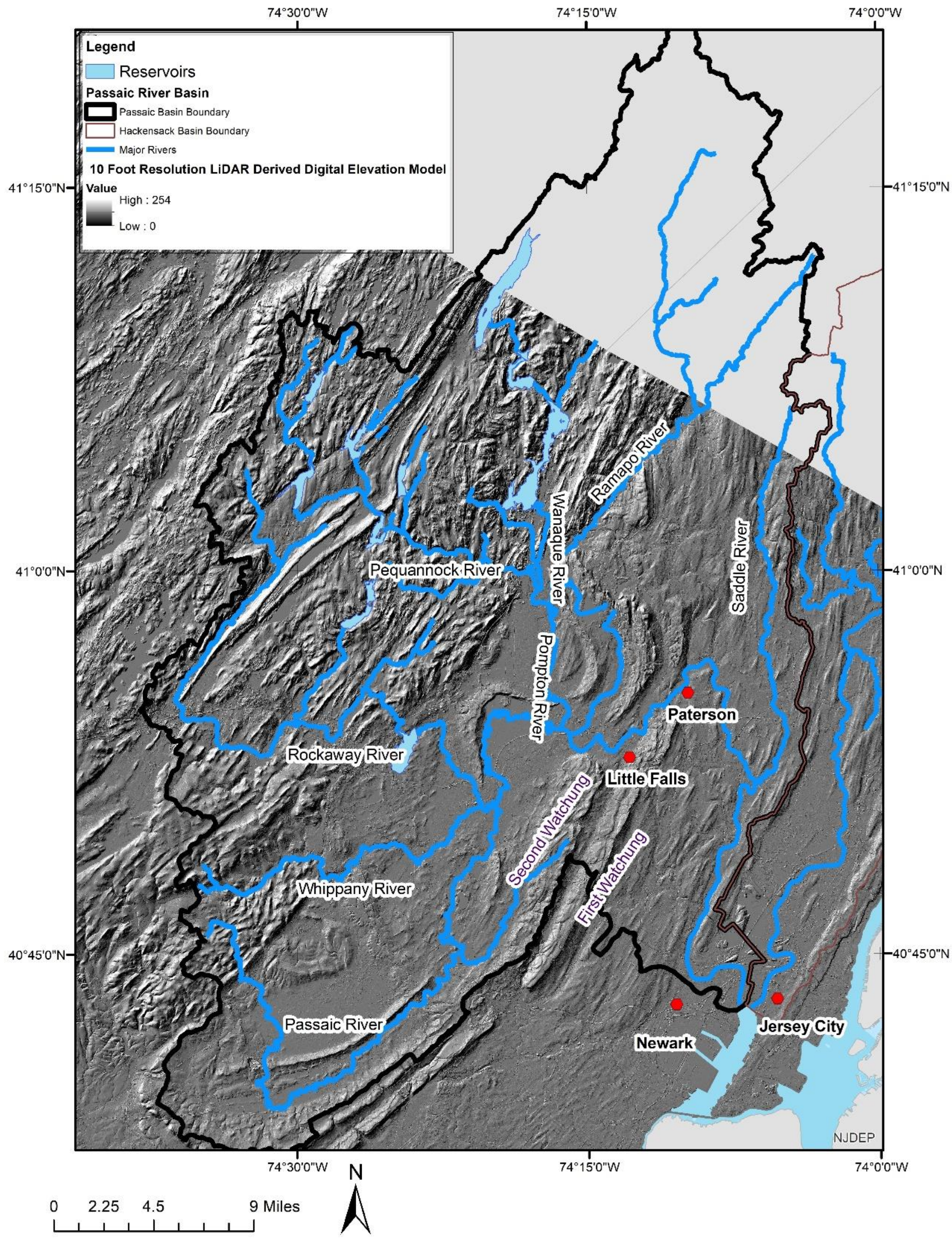


Fig. 8: Lidar elevation imagery of the Passaic River basin showing the contrast between the topography of the Highlands and Central Basin. Data from (USGS, 2016; NJDEP et al., 2010b; NJDEP et al., 2010a).

For this study, discharge data from the selected gages is used to analyze the effects that the contrasting gradients between the Passaic and the Highland tributaries have on flooding. The sections of the rivers along which the gages are located may vary in slope from the average slope of the rivers. Human alterations of the floodplain and natural fluvial processes may also have slightly changed the current average gradients of the rivers since 1902. Both the Passaic River gage at Little Falls and the Pompton River gage at Pompton Plains are located along stretches of river with gradients of about 1.2 x 10^{-3}, which are higher than their average gradients recorded in 1903 (Spitz, 2007). This is likely because of the proximity of the Pompton Plains gage to the Highlands, and the drop of the river through the basalt of the Watchungs at Little Falls. The gages along the tributaries of the Pompton and Passaic range from gradients of 6.6 x 10^{-4}- 2.1 x 10^{-4}, which is significantly lower than their average gradients (Hollister and Leighton, 1903; Spitz, 2007). This is likely due to many of the gages being located at or near dams, reservoirs, and lakes, which are located in areas that are naturally flatter, or have been levelled for construction.

CLIMATE

HOLOCENE CLIMATE

Holocene climate in New Jersey showed variability in wet/dry and warm/cold cycles spanning millennia. Sediment deposits from White Lake in northern New Jersey, about 30 miles east of the Great Piece Meadows, offer insight into the timing and duration of these cycles (Li et al., 2006) . Nine warm/cold ~1500 year climate cycles during the Holocene have been identified from North Atlantic deep-sea cores with ice rafted debris, and correlated with ~1500 year cycles interpreted from ice rafted debris in Northern Atlantic Ocean sediment cores (Bond et al., 2001; Li et al., 2006). The lake level of White Lake is mainly controlled by groundwater level fluctuations, which is strongly affected by climate-controlled moisture abundance (Li et al., 2006). Sediment records of lake levels at White Lake strongly parallel with ice rafting records from the North Atlantic Ocean, showing low lake levels (dry periods) at ~6.1, 4.4, 3.0, and 1.3 ka corresponding with determined cold events. These dates are corroborated by pollen records (Li et al., 2006; Willard et al., 2005). Lake records from nearby lakes, such as Fayetteville Green Lake in New York, also match up with North Atlantic Ocean recorded cold events at ~8.3 and 9.5 ka. Indication of low lakes levels in White Lake at ~4.4 ka and evidence of a low frequency period of storm related floods from 13 New England lakes signify a widespread dry period from ~5- ~3.5 ka (Li et al., 2006; Noren et al., 2002). Higher frequency periods of storm related floods to the New England lakes were determined by large influxes of terrigenous sediment, the greatest peaks being at 11.9, 9.1, 5.8, and 2.6 ka, with the most current increase occurring around 600 years B.P. at the beginning of the Little Ice Age in the early 1400s (Noren et al., 2002; Ramanujan, 2007). These fluctuations in storminess are related to changes in the flow of the Arctic Oscillation, which also may be responsible for climate variability in the Holocene (Noren et al., 2002). Therefore, Holocene periods of storminess in New England may also have extended to Northern New Jersey. A study of cores from Silver Lake in northwestern NJ also gives insight into the climate of the early Holocene (Zelanko et al., 2012). Brief, significant decreases in carbonate levels between 11.1 and 11.3 ka suggest lower lake level and therefore dry conditions. A change in sediment composition from marl to peat and an increase in silica levels also suggests a drier climate, for it indicates lower lake levels and an increase in erosion. Records of high lake levels at Lake Blauvelt in Northern New Jersey indicate wet periods from ~10-8,250 ka B.P. and ~4 ka – 250 year B.P. (Getch et al., 2016).

MODERN CLIMATE

New Jersey has a modified continental climate, meaning it generally has hot summers and cold winters (Paulson et al., 1991). The main sources of moisture for New Jersey are the Gulf of Mexico and the Atlantic Ocean. Winter climate in New Jersey is mainly controlled by an intensified high pressure system located over Central Canada and the north-central US, which sends cold air masses southward, over New Jersey. Occurrences of low temperatures usually follow in the path of a cold front, which itself is usually paired with a strong low pressure system with significant amounts of rain and snow. Winter and early spring storms involve moderately intense, widespread steady rainfall. These storms are prone to greater flooding than summer storms (Paulson et al., 1991). In the winter, frozen ground decreases soil permeability, while existing snow cover melts during rainstorms, increasing storm runoff. The main system affecting New Jersey in the summer is the Bermuda High, which is a semipermanent high-pressure cell reaching to eastern Gulf of Mexico (Paulson et al., 1991). Humid air is brought up from the Gulf of Mexico, resulting in hot and muggy summers. Changes in air flow and passing fronts can result in thunderstorms with intense rainfalls, which can cause locally severe floods. Climate varies in different parts of the state. The northern part of the state is affected by storms across the Great Lakes Region, while most of southern New Jersey is not (Paulson et al., 1991). Coastal storms originating from or around the tropics cause heavy rainfall and winds near the coast, although storms can move inland and up north as well. Annual precipitation usually ranges from 40-52 inches, and peaks slightly in the summer and winter along the coast (Paulson et al., 1991). Annual snowfall ranges between 13 inches (in the south) and 50 inches (in the northern highlands). Percentage of annual precipitation from tropical storms is 4% in the north and 6% in the south, although tropical storms are responsible for 30% of the rainfall in September (Paulson et al., 1991). Overall, New Jersey has mostly mild weather and ample precipitation, with occasional extreme storms.

Droughts

Five major droughts have occurred in New Jersey during the period of time that this study encompasses. These droughts periods are 1929-1932, 1949-1950, 1953-1955, 1961-1966, and 1980-1981, 1999-200, and 2002 as well as a moderate drought during 1984-1985 (Bauersfeld and Schopp, 1991; Hoffman and Domber, 2004; Office of the New Jersey State Climatologist, 2002; Paulson et al., 1991). The most intense drought lasted five years, occurring from 1961-1966. The recurrence of a drought of this degree was 25-50 years in most of New Jersey, but over 50 years in northern New Jersey, where streamflows deficiencies were greatest. The drought of 1929-1932 was a regional scale drought that affected most of the Northeast, as well as the second most severe drought in New Jersey's history (Paulson et al., 1991). The 1949-1950 drought was less extreme than the 1929-1932 drought, but included the driest June on record in the history of New Jersey stream gages. This drought had the greatest impact on northeast NJ, where a drought of that magnitude had a recurrence interval of more than 10 years (Paulson et al., 1991). A drought as severe as the one in 1953-1955 was slightly rarer, with a recurrence of 15 years. The 1984-1985 drought varied in severity across the state, with recurrence rates of 4-9 years in the northern part of central NJ and the southwest. However, northern and eastern areas of New Jersey had a recurrence interval ranging 10-20 years, with the 1984-1985 drought being severe enough for a state of emergency to be declared for most of northeastern NJ. The drought of 1980-1981, which was mostly statewide, also had a recurrence interval in this range, from about 10-25 years, and caused the Boonton Reservoir in the Passaic River Drainage Basin to have record low water levels at the end of January, 1981 (Paulson et al., 1991). Drought emergencies were also declared in Northern and Northeastern NJ in 1995, 1999, and 2001-2002 (Hoffman and Domber, 2004; Office of the New Jersey State Climatologist, 2002)

HUMAN SETTLEMENT AND FLOODING PROBLEMS

New Jersey's Passaic River Basin has been plagued by chronic flooding problems during times of increased rainfall and snow-melt since colonial times, causing 13 federal disaster declarations in NJ since 1903 (US Army Corps of Engineers; US Army Corps of Engineers, 1987; US Army Corps of Engineers, 2017). Flooding and contamination of river sediments within the Passaic River basin has been a persistent problem for over a century (Brydon, 1974). Recent major floods within the basin have occurred in 2011, 2010, 2007, 2005, 1999, 1989, 1992, 1987, 1984, 1983, 1979, 1978, 1977, 1975, 1973, 1972, 1971, 1968, 1955, 1945, 1936, 1903, 1902, 1896, 1882, 1878 and 1810 (US Army Corps of Engineers; US Army Corps of Engineers, 1987; US Army Corps of Engineers, 2017). Because much of the areas along the river are urbanized, chronic flooding has led to major property damage, and contamination of river sediments by pollutants

EFFECTS OF URBANIZATION ON WATERSHEDS

Urbanization and land development can have a profound effect on the both the hydrology of the entire drainage basin and the discharge and channel morphology of individual rivers. One aspect of urbanization that affects hydrology is the creation of impervious surfaces such as concrete and asphalt roads, parking lots, and airports. These impervious surfaces cause the surrounding area to decrease in infiltration, thereby increasing runoff within the drainage basin and subsequently increasing river discharge (Dunne and Leopold, 1978). Removal of vegetation, which contributes to soil stability and absorbs rainwater, also contributes to increased runoff. Increased runoff also reduces groundwater recharge and baseflow discharge in urban streams, though baseflow may also be enhanced by urban wastewater effluent (Hirsch et al., 1990; Paul and Meyer, 2001). Impervious surface cover downstream in a catchment system drains flooding faster than stormflow from a forested area further upstream (Paul and Meyer, 2001). Peak river discharges are also higher in urban catchment areas (Leopold, 1968). However, flood discharges with reoccurrence intervals that span over a longer period of time are less likely to be affected by urbanization than discharges of frequent flooding (Espey et al., 1965; Hirsch et al., 1990). Peak flood discharges may also be lowered due to water detainment by blocked culverts, drains, ditches, or other urban features (Hirsch et al., 1990).

Sediment supply and bankfull discharge also increase because of construction, which loosens and exposes sediment that can be easily washed into rivers. This increase in sediment leads to the aggradation within the river (Paul and Meyer, 2001). Sediment fills the river channel, decreasing channel depth and contributing to greater floods and overbank deposits (Paul and Meyer, 2001; Wolman, 1967). During this phase the width to depth ratio of the channel either increases or remains constant. Although construction originally increases sediment supply and causes channel infill, the resulting effects of urbanization ultimately leads to channel incision. Erosion results from the increased impervious surface cover in urban environments (Arnold et al., 1982; Booth and Jackson, 1997; Dunne and Leopold, 1978; Leopold, 1968). The resulting increase in bankfull floods and stream discharge causes channel incision to occur (Paul and Meyer, 2001). Channel incision is dependent upon sediment influx and the amount of sediment that can be carried by stream flow. If the amount of transported sediment is greater than the influx of sediment, downcutting will occur, and the channel will be enlarged (Booth, 1990). After incision, channels begin to migrate laterally and bank erosion and channel widening begins (Booth, 1990; Booth and Jackson, 1997; Trimble, 1997).

SETTLEMENT AND LAND USE IN THE PASSAIC RIVER BASIN

History

The Passaic River has experienced major flooding since colonial times (Anderson et al., 1999). Artifacts recovered around the vicinity of Little Falls, Paterson, Two Bridges, and Chatham determined that the Lenape lived in this area before the arrival of Europeans in the 1600s (Brydon, 1974). One of the first European settlements was established in Morris County spurred by the discovery of iron ore deposits and the use of the Passaic for waterpower. Wood-burning forges became established along the

Whippany River, producing large amounts of charcoal (Brydon, 1974). Pompton Plains and Lincoln Park were settled during the early 1700s. Dutch settlements began to spread during this time along the joining of the Whippany and the Rockaway Rivers, near Parsippany-Troy. Mills established along the Passaic, which were kept running until the late 1800s and early 1900s. Wetlands such as the Great Piece Meadows were used for hay farming.

By 1770, flooding issues as a result of dams were already apparent. The river was dammed at the head of Little Falls to provide water power for the mills of Capt. James Grey (Brydon, 1974). When the river rose, the dam caused the river to overflow the upstream banks and flood the surrounding farmlands, causing crop losses. Action was soon taken to combat this problem. In 1772, the General Assembly passed an Act for the clearing and dredging of the Passaic River. In 1774, Deepavaal Ditch, or Mill Ditch, was formed to alleviate flooding by promoting drainage within the Great Piece Meadows (Brydon, 1974). Despite continuing problems, hay was grown in the Great Piece Meadows until the 1930s, and nearby land was farmed to supply goods to New York City (Anderson et al., 1999; Brydon, 1974)

In 1790, a 4.5 foot high dam was built above Great Falls (8-10 miles downriver from the study site), which supplied 3 canals that provided water power to mills. This was regulated by guard gates and lowered the Morris Canal's water volume, which was noticeable during times of drought when the Falls were reduced to mere trickles. Dam size and diverted waters increased over the years, as did pollution (Brydon, 1974). Legislative approval was given to owners of flooded lands along the Passaic to break up rock reefs within the river.

With the help of Alexander Hamilton, Governor Paterson signed a charter for the Society for Establishing Useful Manufactures in 1791 (Brydon, 1974). This lead to the Paterson at Great Falls becoming the nation's first industrial city (Anderson et al., 1999). Water power from the Passaic River and Great Falls was used to power mills that manufactured silk, train workings, and Colt guns. A small dam was built above the Great Falls, which was then converted into a reservoir (Brydon, 1974). Water was taken from the reservoir 150 feet to the factory. Due to the War of 1812, the city of Paterson once again experienced a boom in manufacturing in 1814. In the 1800s, railroads became prominent and allowed easy access into New York City (Anderson et al., 1999). With the invention of the automobile, extensive highways and road systems were developed, including the building of the George Washington Bridge in 1931. The Morris Canal, which was constructed to carry coal from Lehigh Valley, Pennsylvania to the New York Harbor, was completed in 1831 and extended to Jersey City in 1836 (Brydon, 1974; Rice, 2007). In 1820 and 1834, legislation was passed for removal of river obstructions in the lower Passaic (Brydon, 1974).

The Dundee Company added 6 feet to the dam at Paterson in 1833 and extended it across the river, dug the first canal (Brydon, 1974). The dam was partly swept away, but a new dam was built at Great Falls around 1838. By 1844, the Great Piece Meadows was producing 1,500 tons of hay and Little Piece Meadows were producing 500 tons of hay. The traprock reef and the dam at Little Falls were said to cause more frequent and prolonged flooding that submerged the hay. Organized efforts to remove blockages were initiated, and a 2,000 ft^3 section of rock as removed, allowing water to drain more easily (stagnant water did not remain as long), and the area flooded less (Brydon, 1974).

Deepavaal Ditch was widened and deepened in 1858. However, flooding problems continued. In 1868, Mill owners, who built a dam below the rock reef at Little Falls prevented further lowering of the reef and received authorization to repair and extend the dam. This caused more flooding problems than the trap rock reef. Because of this, Passaic Drainage Commissioners from 1889-1890 contracted Alfred B. Nelson to blast away trap rock above the dam at Little Falls, reducing the main fall below the dam, but work was not completed (Howell, 1898). 3,000 yards of rock were left blasted but not removed form Little Falls (Cook and Smock, 1887). A Contract with Morris and Cumings Dredging Company of NYC to remove rock at Little Falls, but work suspended due to lack of funds (Howell, 1898). The US Geological Survey concludes many of the early mills were still in operation in 1894 in Chatham and Stanley (Brydon, 1974). Maintenance of Deepavaal Ditch provide jobs for depression-era workers in the 1930s, which may have helped with drainage in the Central Basin. The Morris Canal was officially abandoned in 1929 (Rice, 2007). Within the past 50 years, an increase in housing, roadways, and structures such as the Willowbrook Mall have increased impermeable surfaces and encroached on the Passaic's floodplain, leading to more destructive floods (Youssef, 2014) (Fig. 9).

Development and Population

Because of the rapid spread of industry in Paterson, Jersey City, and Newark, population around the Passaic River Basin grew rapidly. In 1860 Newark's population was already above 75,000 (Brydon, 1974). Population densities for Passaic County have gone from 1,300-3,000 people per square mile in 1930, to 3,500-6,200 in 2007 (Siegel et al., 2009). Morris County has gone from 50-600 people per square miles in 1930, to 650-1,200 in 2007. Essex county has remained constant from 1930-2007 with 6,500-7,400 people per square mile.

While Hatfield Swamp, the Great Piece Meadows, and wetlands along the floodplains of the Wanaque, Pequannock, Pompton, and Ramapo Rivers have been left undeveloped, urban and suburban build up has occupied most of the Passaic's floodplains (Fig. 10). Drainage of the Passaic Basin's wetlands, including the Great Piece Meadows, for farming and mosquito control has also decreased their ability to retain water, thereby increasing the this discharge and damage of floodwaters (Brydon, 1974). It is estimated that Passaic County lost about 66% of its wetlands from early colonial times to the mid-1970s (US Army Corps of Engineers, 1987). Between 1940 to the mid-1970s, Passaic County lost 853 acres of wetlands, amounting to 14% of its existing wetlands from 1940. Wetlands within the Central basin decreased from 29,308 acres in 1902 (Hollister and Leighton, 1903) to about 24,485 acres in 1984 (US Army Corps of Engineers, 1987). From 1948-1978, the Passaic River Basin lost close to half of its wetlands (Branch, 1994). In 1978 there were about 268,930 developed acres in the basin (US Army Corps of Engineers, 1987), but between 1985 and 2007, another 900 acres of forested land was developed (Martin and others, 2011). For the entire Passaic River Basin, 4.5% of wetlands was lost between 1985 and 2007 (Martin and others, 2011).

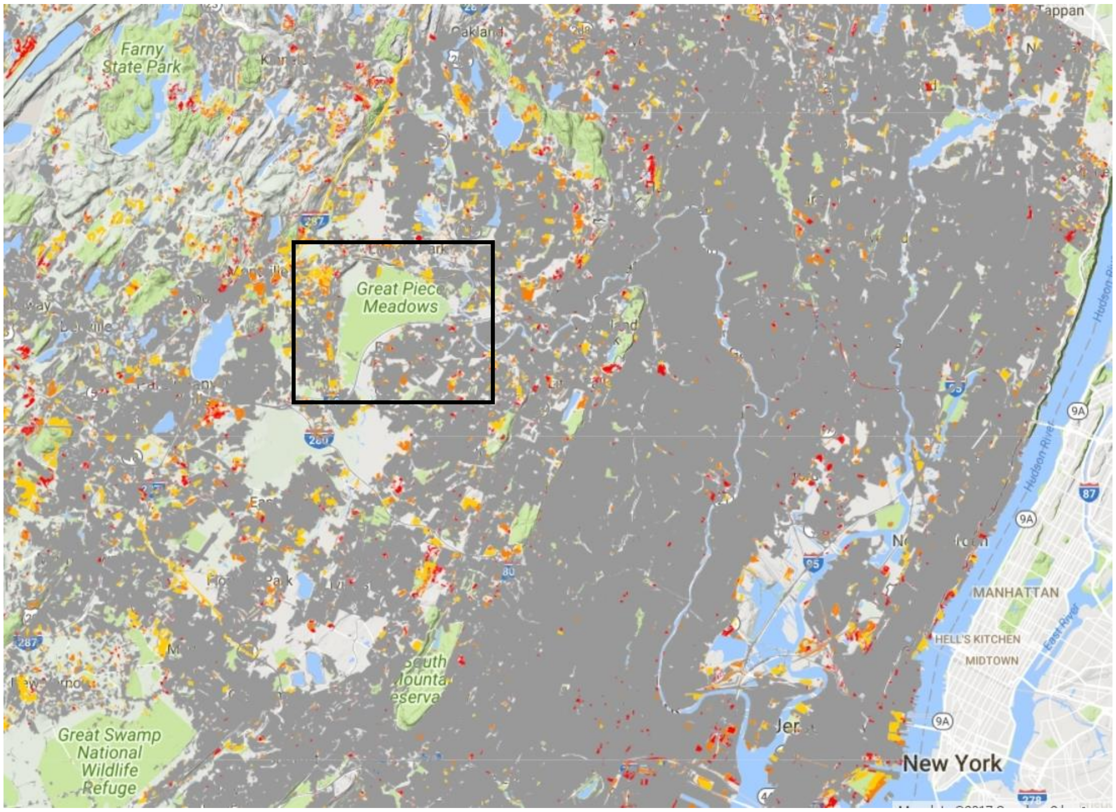


Fig. 9: Map of land use changes in northern New Jersey. Grey indicates areas urbanized before 1986. Yellow are areas urbanized between 1986-1995. Orange indicates urbanization between 1995-2002, and red indicates areas urbanized between 2002-2007. Study area outlined with black a box. Map is from Rowan's geospatial research lab, using data provided by the NJ Department of Environmental Protection (Rowan University).

STUDY SITE DESCRIPTION

THE GREAT PIECE MEADOWS

The Great Piece Meadows is a large, forested wetland that makes up the floodplain of the Central Passaic River Basin. At its current boundaries above Rt. 80, the Great Piece Meadows covers 2,343 acres along the river as it forms the northern border between Essex and Morris Counties, covering Fairfield Township, Lincoln Park, and Montville, and ends at Two Bridges, where the Pompton River joins the Passaic River (Cavallo et al., 1994). Large marshland areas within the Passaic River Basin, including the Great Swamp, Troy Meadows, and Hatfield Swamp, which lies just upstream of the Great Piece Meadows, overlie lake deposits from Glacial Lake Passaic and the postglacial lake stages that followed after the recession of the ice (Hoffman, 1989; Stanford, 2007b) (Fig. 10). Hatfield Swamp lies along the Passaic for nearly two miles, occupying an area of about 600 acres in Roseland and West Caldwell in Essex County, and encompasses the location at which the Whippany and Rockaway Rivers join the Passaic (Brydon, 1974) (Fig. 10). Widespread drainage of this area has begun to alter it from a swamp to a woodland environment, but prior to drainage, the area was prone to widespread flooding during increased rainfall.

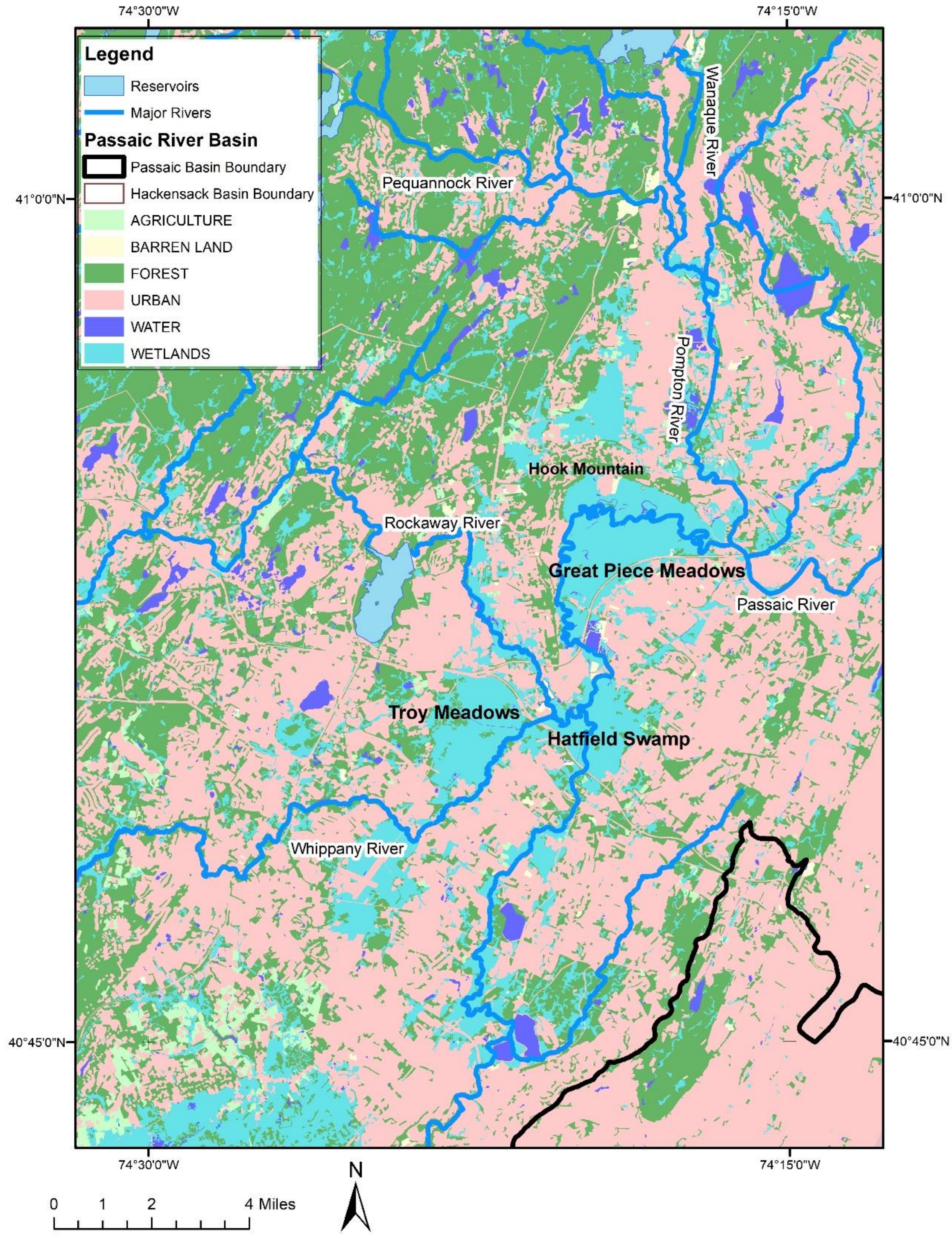


Fig. 10: 1986 Land Use and wetlands within the Central Passaic Basin. Data from (NJDEP, 1998; NJDEP et al., 2010b)

The Great Piece Meadows (also broken into the Great and Little Piece Meadows) was a large, dense, tree-filled swamp when European colonists first settled the area (US Army Corps of Engineers, 1987). The wetland environment and frequent inundation deterred colonist from building houses and villages in the Great Piece Meadows, but most of the dense forestation was later removed as the land was cleared for meadow hay farming to use as feed for livestock (Brydon, 1974). Crude drainage ditches were also dug to prepare the marshland for hay agriculture (US Army Corps of Engineers, 1987). When its current owner at the time of the Revolution, a British Loyalist, fled to England, the township divided the lands and sold them to local farmers. Frequent flooding of the area and the spread of disease by mosquitoes hindered hay farming. This lead to the creation of Deepavaal Brook in the 1770's to divert floodwaters, and the excavation of drainage ditches throughout the years for mosquito control (Brydon, 1974). As horses and cattle-driven plows and cars were replaced by tractors and cars during the early 20th century, the need for hay lessened, and the crops in the Great Piece Meadows were left unharvested (Brydon, 1974). Hay farming ceased completely around 1930, as local subsidence caused by the shrinkage and decomposition of peat and the infilling of drainage ditches contributed to the soil being too saturated for farming (Fig. 11). The end of farming allowed the wetland to transition back to a forested/shrub ground cover stage, although this transition is hindered in flat or backswamp areas with poor drainage (US Army Corps of Engineers, 1987).

Fig. 11: 1930's aerial imagery of the Great Piece Meadows, with core locations of primary cores Oxbow2, RC2, and TZS2 as red circles. The yellow circle is the site of OX1, a test core. Data from (NJDEP, 2005)

After its use as hay field was discontinued, work began to preserve the Great Piece Meadows for its natural flood storage and its potential in reducing the impact of high magnitude floods downstream. Wildlife Preserves, founded in 1952, bought land in the Great Piece Meadows for preservation, later joining with the DEP in 1981 to protect and preserve the Great Piece Meadows (Wildlife Preserves, 2012-2015). In 2003, the US Army Corps of Engineers acquired the Great Piece Meadows for NJ DEP's Passaic River Preservation of Natural Flood Storage Areas project. New Jersey Natural Lands Trust, Wildlife Preserves, NJDEP, Fairfield Conservation and Sportsmans Association, as well as several landowners in Fairfield now co-own and co-manage The Great Piece Meadows. The Great Piece Meadows are currently closed to motor vehicles, so the upper layers of soil remain relatively undisturbed (Wildlife Preserves, 2012-2015).

OXBOWS

Oxbows within the Great Piece Meadows were chosen for coring because the Great Piece Meadows is the largest undeveloped wetland in both the Central Basin and Lower Valley of the Passaic River Basin, and because of its influence on flooding downstream. The bulk of the damage done by the Passaic's flooding is experienced downstream in the Lower Passaic. While there is historical data from the gage at Two Bridges at the edge of the Great Piece Meadows, a longer, more complete record is found at the Little Falls gage near Paterson, making flooding and hydrologic data from Little Falls integral to this study. A coring site within the lower Passaic River basin may seem more logical to compare past flood deposits with documented flood discharges, but the Great Piece Meadows proved to be a better site for several reasons. One main factor is that almost no areas of undeveloped floodplain exist along the lower Passaic, making it difficult to find a coring site that retains current flood deposits as well as maintains an uninterrupted record of past flood deposits. The Great Piece Meadows is currently protected from urban development and has resisted major buildup since it was first settled, leaving it mainly undisturbed. While it is possible that hay farming, which lasted from the early 1700s to the 1930s, disturbed the uppermost layer of soil, disturbances from hay farming would likely have been minor, especially with unfavorable farming conditions due to flooding. Also, it is likely that the oxbows remained undisturbed by farming because they would frequently contain standing water. Strong evidence for this is seen in detailed historical maps dating back to the late 1880's, in which the two northern oxbows in Lincoln Park are shown in the map symbol for water bearing channels, and in 1930's aerial imagery, which shows agricultural fields in the majority of the meadows, but wooded, unfarmed areas around the oxbows (Fig. 11).

The Great Piece Meadows was also chosen because its abandoned oxbow meanders provide coring sites for studying the record of flooding deposition. While drainage for farming and mosquito control have reduced the oxbows from a lake environment to a wetland environment, the flat topography and their connection to the main channel cause them to become inundated during flooding. In their 1987 report, the US Army Corps of Engineers labeled the oxbow areas as "seasonally flooded", (US Army Corps of Engineers, 1987). Flood inundation modeling of the Passaic River at Pine Brook, NJ by NOAA also show the three oxbow coring sites as being inundated at minor flooding stage (gage height of 18 ft) (National Weather Service et al., 2016). Though the oxbows can longer be classified as a lake depositional environment, they are often filled or partially filled with standing water. In the oxbows on the Lincoln Park side of the floodplain, both oxbows contained standing water in some areas, and the water table was at or near the surface at the coring sites. The oxbow on side of the floodplain bordering Hook Mountain did not contain standing water, but the water table was reached about a foot below the surface. The fact that they become submerged during flooding and their ability to retain water make the oxbows the best location for preserving a record of flood deposited sediments along the floodplain.

The Great Piece Meadows also proved to be the prime location for studying flooding within the Central Basin and Lower Valley because of the regional geology and its position along the Passaic River. The Great Piece Meadows is situated between three of the Passaic's main tributaries; the Whippany and the Rockaway Rivers connecting upstream at the south end of the Meadows, and the Pompton connecting downstream at the eastern edge of the Meadows. These tributaries carry water from rains and snowmelt in higher elevations from the New Jersey Highlands, which are a main factor in major flooding along the Passaic (Hollister and Leighton, 1903). After joining with the Pompton River, the Passaic continues downstream to flow over the Second Watchung Mountain at Little Falls, about 4 km from the Great Piece Meadows. This narrow outlet within the basalt of the Watchung Mountains constricts the waters of the Passaic during heavy flooding, causing floodwaters to back up and become retained within the Great Piece Meadows (Brydon, 1974; US Army Corps of Engineers, 1987). This issue was recognized in the early 1800s, and legislation was enacted in 1844 to remove 2,000 cubic feet of rock from the river channel (Brydon, 1974). After 1872, another effort was made to widen this outlet, resulting in the destruction of the Little Falls waterfall and the removal of hundreds of tons of rock. Although this helped to alleviate some of the severity of flooding, floodwaters are still impeded at Little Falls, causing them to back up and be detained within the Great Piece Meadows, giving a direct correlation between flood discharge measurements at the Little Falls stream gage and a record of flood deposits within the Great Piece Meadows.

METHODS

FIELD

Coring

Cores were taken from three meander cutoffs using a Livingston drive rod piston corer. This type of corer can be used in water 30 meters deep or a wetland environment to take a succession of cores up to one meter in length. The corer is comprised of a 1.27 meter barrel containing a piston held by a cable (Myrbo and Wright, 2005). A metal rod and handle are attached to the barrel in order to remove it from the ground and push the core out of the barrel. A team of at least three people was needed to extract the core. Three complete cores, labeled Oxbow2, RC2, and TZS2, measuring each about two meters in length, were taken at three different meander cutoffs along the Central Passaic River within the Great Piece Meadows (Fig. 12). Each section of core was encased in plastic tubing and wrapped in plastic wrap. These meander cutoffs, formally oxbow lakes, vary seasonally in saturation and inundation but still remain a wetland environment.

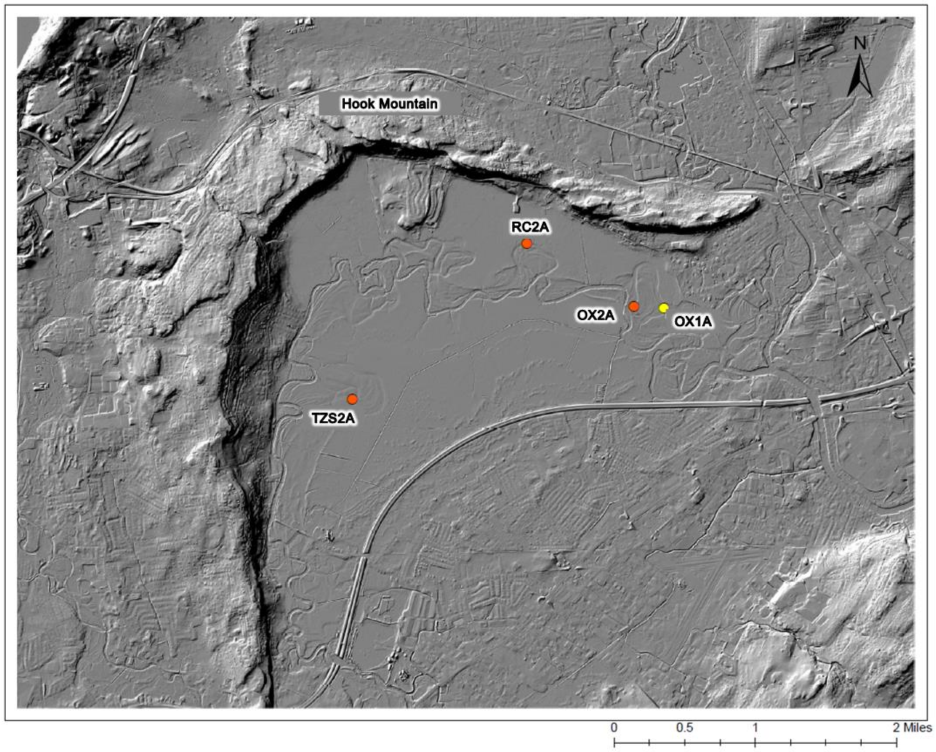
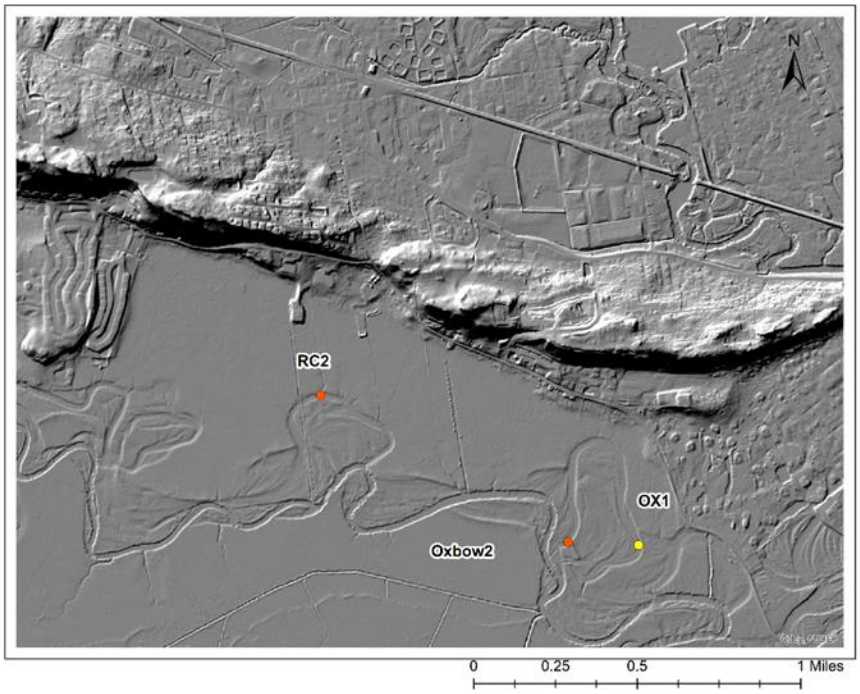


Fig. 12: Lidar imagery (top) and 2007 aerial imagery (bottom) of the Great Piece Meadows, primary coring sites (red circle) and test core sites (yellow). Data from (OGIS, 2012--2013; USGS, 2016)

Core Oxbow 2 The first oxbow was cored in November of 2014 (Fig. 13). Despite being a wetland environment, the oxbow was fairly dry during the month of November, and only contained a few areas of shallow standing water. The water table was high throughout the oxbow. A preliminary core was taken on the downstream side of the site of the final Oxbow 2 core site to test out the corer and determine the best area to retrieve a complete core. While the first core site was abandoned due to difficulty in extracting the core and an overabundance of surface vegetation, the test core was labeled as "OX1" and kept for additional analysis. The Oxbow2 core was taken in four intervals: Oxbow2A, Oxbow2B, Oxbow2C, and Oxbow2D, combining to make a complete core 212.5 cm in length.

Core RC2 The second oxbow was cored in May of 2015 (Fig. 13). The wetlands of the Great Piece Meadows were more saturated during late spring than they were in the fall, as shown by an increase of areas with shallow standing water and marshy conditions within the oxbow. Two initial cores, RC1A and RC1B were taken about 30 feet away from the final core site, RC2. These two cores were saved for further analysis after the original borehole was abandoned due to instability. Core RC was cored in three sections, RC2A, RC2B, and RC2C, with a combined length of 213 cm.



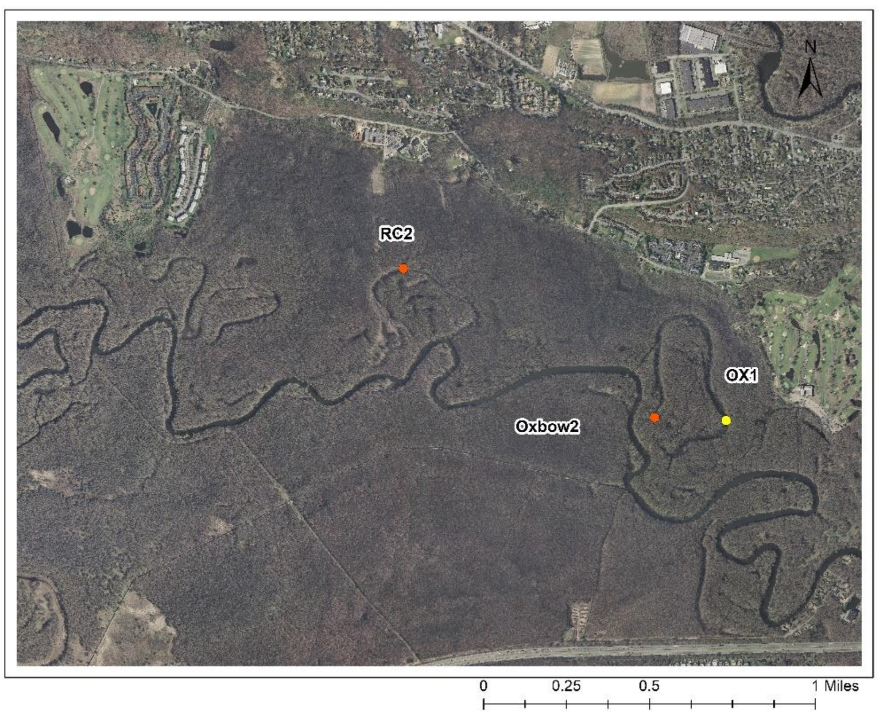


Fig. 13: Lidar (top) and 2007 aerial imagery (bottom) of Oxbow2 and RC2 coring sites. Data from (OGIS, 2012--2013; USGS, 2016)

<u>Core TZS2</u> The last oxbow is located on the opposite side of the river from the first two oxbows was cored. (Fig. 14). The site was surveyed using a bucket auger prior to coring in order to determine the best coring location within the oxbow. The water table was reached less than a foot below the surface. The first auger location, situated near the coring site, was augered down to about 8 feet, with a thick sand layer encountered around 5 feet down. A second auger site, located at the northern end of the oxbow (Fig. 14), was drilled down to 9 feet. A sand layer similar to that at the first site was encountered closer to the surface than the previous site, and clay was sampled from the bottom of the auger hole.

Two cores, TZS1 and TZS2 were initially taken within 20 feet of each other, but only TZS2 was cored to completion because of loss of the core at during extraction at TZS1. TZS1 was saved for analysis. TZS2 was cored to a total of about 216 cm and is comprised of 5 sections: TZS2A, TZS2B, TZS2C, TZS2D, and TZS2E. Only the first four cores, with a total length of 168 cm, are considered to contain river and floodplain sediments of the Passaic. During the coring of the final section, TZS2E, the corer encountered a thick layer of sand. Since the Livingston corer is designed to core wetlands and lakes, the compacted sand prevented the piston from pushing out the complete core from the barrel, most of the core had to be scooped from the barrel by hand. Because of this, the sediments of core TZS2E is not in exact stratigraphic order, and some of the core may have been lost. However, this large layer of sand does not appear to contain floodplain deposits, so while it contributes to the study of the overall surficial stratigraphy, it does not impact the study of the Passaic River's flooding.

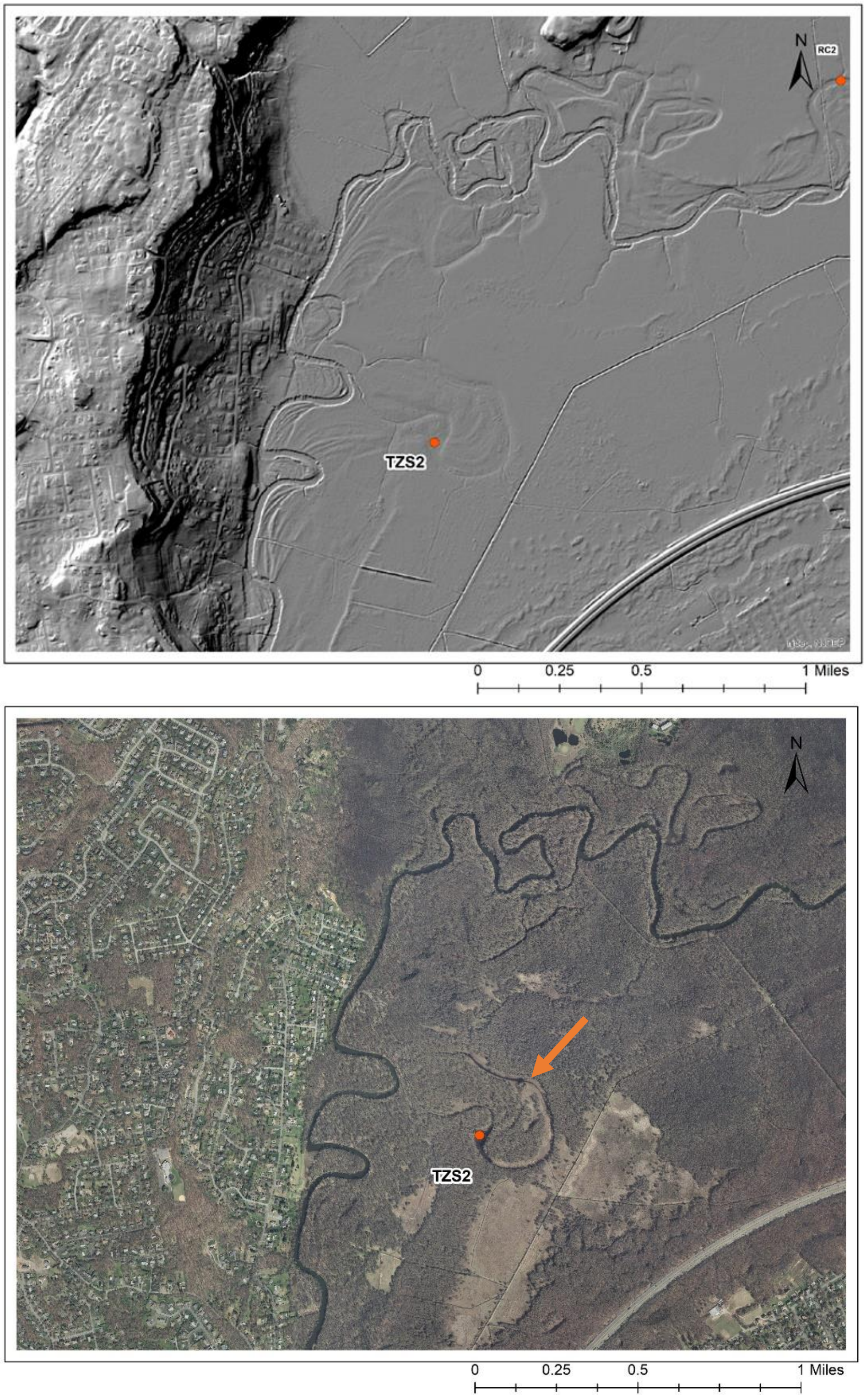


Fig. 14: Lidar (top) and 2007 aerial imagery (bottom) of the TZS 2 coring site. The second auger site is indicated by the orange arrow. Data from (OGIS, 2012--2013; USGS, 2016)

LABORATORY

Grain Size

Grain size was determined using the Malvern particle analyzer. Wet sieve and pipette analysis was used for grain size analysis for samples that contained grains too large to pass through the Malvern.

Wet Sieve and Pipette Analysis Grain size was determined using the standard pipette grain size analytical method (Folk, 1980). Each sample except for Sample 7 from core RC2C was weighed to about 10-30 grams. The smaller sized samples (<20 grams) were from core TZS and RC, since the 2 cm diameter core could not provide several 30 gram samples of sediment without being mostly destroyed. The samples were placed into a clean beaker and diluted with distilled water. The samples were stirred with a metal sampling spoon to disaggregate the samples. Between 2-5 ml of 30% hydrogen peroxide were added to the samples to remove organics, which were covered and allowed to sit for a couple days. The samples were washed with distilled water and stirred to create a slurry. This slurry was poured through a series of sieves and washed with distilled water, separating the samples into the 1.0, 0.5, 0.25, 0.125, 0.0625, and 0.0039 mm fractions following the grade scale of (Wentworth, 1922). The covered sieves were allowed to completely air dry and were then weighed to 0.01 g. The sieved fraction was calculated against the total sample weight to get the percent of each grain size within the sample.

For the majority of the samples, only sand was analyzed. Only samples 1 and 4 had further analysis for silt and clay ratios. Silt and clay ratios were determined using the pipette method. Silt and clay that passed through the 0.0625 mm sieve were collected in a 1000 ml beaker. The beaker was filled to the 1000 ml mark with distilled water. One gram of sodium hexametaphosphate was added per 1000 ml of water. After the addition of the deflocculant, the mixture was stirred thoroughly with a metal sampling spoon and left to sit overnight. A 1000 ml beaker filled with tap water was also allowed to sit overnight to monitor water temperature.

For pipette analysis, the sample was placed in a 1000 ml cylinder and filled to the 1000 ml mark with distilled water. Two pre-weighed 50 ml beakers were labeled for "silt and clay" and "clay". The sediment in the cylinder was then agitated into suspension with a plastic stirring rod. Using a 20 ml glass pipette, samples were taken at times and depths corresponding to Stoke's Law of settling. After thoroughly stirring the sample to make sure all sediment is within suspension, the pipette was lowered to 20 cm within the cylinder. After 20 seconds, a 20 ml sample was extracted and emptied into the 50 ml beaker labeled "silt and clay". At 2 hours and 3 minutes from the start time, the pipette was lowered to 10 cm, and a 20 ml sample was extracted and emptied into the beaker labeled "clay". After both samples are collected, the beakers were placed in an oven to dry at about 65°C for at least 24 hours. The dry samples were placed in a dessicator to cool and then weighed. The silt/clay amount was determined by subtracting the weight of the deflocculant and multiplying that amount by 50.

<u>Malvern Sample Preparation</u> Grain size analysis was primarily done using the Malvern Mastersizer 3000 laser diffraction particle size analyzer at the Institute of Marine and Coastal Sciences at Rutgers University. Measurements are taken by passing a laser beam through a dispersed particulate sample and analyzing the scattering angles and

intensities as the laser interacts with the particles. Larger particles scatter light at lower angles than small particles, so particle size can be inferred relative to the angular variation of the laser beam. This data is processed using the Mie theory of light scattering in order to determine the particle size as a volume equivalent sphere diameter (Malvern Instruments Ltd). The results for each sample is the average of three separate measurements during each run. Particle size distribution for a sample is given in percent of volume density. Sediment classification (sand, silt, clay) is given in percent by volume.

Samples for grain size analysis were taken from each core at locations where grain size boundaries were perceived during the initial logging of the core. Roughly 1 cubic centimeter of sample was obtained from the center of the core, so as to avoid possible contamination from the outside of the core. The samples were placed in plastic test tubes and washed with 30 ml of 30% hydrogen peroxide solution. The samples were then allowed to sit for 24-72 hours before placing them in a hot water bath. They then remained in a hot water bath, set at about 55°C, for 2-3 weeks. The samples were stirred daily and refilled with hydrogen peroxide up the 30 ml mark as needed while in the hot water bath to ensure that the reaction continued. Once the samples have finished reacting, the test tubes were filled to 40 ml with deionized water and capped. The tubes were placed in a centrifuge and run at 3600-4000 rpm for about 20 minutes at a time. Hydrogen peroxide and water were then poured off, and the tubes were refilled to be centrifuged again. Each tube was centrifuged three times to ensure the removal of the hydrogen peroxide. After the final run on the centrifuge, the sample tubes were decanted down to about 5 ml. They were then filled up with 5-7 ml of deflocculant solution (5.5 g sodium metaphosphate/liter). The samples were then stirred with a metal spatula and left to sit for about 48 hours.

Each sample was run through the Malvern at least three times to ensure reproducibility. The sample solutions were stirred vigorously to evenly disperse all particles into the water column. Malvern samples were removed from the solution using a plastic pipette and released into the wet sample dispersion unit, where it was sonicated before measurement. The results for the sample is the average of three separate measurements during each run. Particle size distribution for a sample is given in percent of volume density (Appendix 3, Table 1). Sediment classification (sand, silt, clay) is given in percent by volume.

Radiocarbon Dating

Organics found within the cores were dated using the Accelerator Mass Spectrometry (AMS) technique. AMS radiocarbon dating for all samples except D-AMS 019864 were performed by Beta Analytic. The remaining sample was analyzed by DirectAMS. Dates measured in radiocarbon years before present (B.P.), which is 1950. AMS radiocarbon dating measures sample the sample C14/C13 ratio in comparison to the C14/C13 ratio in Oxacilic Acid II and corrected for total fractionation using machine graphite δ13C. Samples are pretreated to minimize error. Carbonates are eliminated from the sample by a hot HCl acid wash, which is followed by an alkali wash to eliminate secondary organics such as rootlets and additional sediments (BetaAnalytic, 2017). Because of pretreatment, different organics have certain weight requirements to ensure that enough dateable material is present. Organics found within the cores consist of wood, leaves, charcoal, and seeds. Wood, which was the most abundant type of organic material found, requires 3-100 mg for dating. Seeds also require 3-100 mg, but 10-20 mg is recommended. Seeds are considered to provide more reliable dates for the time of deposition because they grow and are released by plants within the same year (Beta Analytic, personal communication). However, it is not guaranteed that seeds occur in a sediment layer contemporary with their deposition, since they could have washed in from other locations or infiltrated lower sediment deposits.

The cores were sampled using a metal scoop, and samples were placed under a microscope in order to extract seeds for dating. Although seeds are ideal for dating, they were not abundant in oxbow sediments. Seed samples that were under 10 mg were supplemented with organic samples of a different material. Wood is subject to the adsorption process by humic and fulvic acids in the soil, which can affect the measured radiocarbon age result. Dead wood that no longer exchanges carbon with the atmosphere may remain attached to the tree for years before it is deposited to the sedimentary record, also affecting the radiocarbon age (BetaAnalytic, 2017).

Mercury

Atmospheric mercury deposition from coal burning accumulates in surficial sediments. Mercury concentrations within the sediment can therefore be used to determine the onset of industrialization and the use of coal powered factories within the surrounding region. The closest major source of atmospheric mercury by coal burning is likely the city of Paterson, located about 8 miles downriver from the core sites (Brydon, 1974) (Fig. 1). Paterson became the nation’s first planned industrial city in 1791, using the waters of the Passaic to power its mills and factories (Paterson National Historical Parks, 2015). While some coal usage was present at the beginning of the 1800s, coal became widely used as a primary source of energy for industry in northern New Jersey in 1830-1840. During this time, new smelting techniques using anthracite coal made steam motors and engines more popular in factories (Clark, 1916). Steam boat passenger services and railroads became established as regular forms of transportation, contributing to coal burning emissions (Brydon, 1974; Rutgers University Libraries, 2011). New developments in transportation also made coal more accessible and more widely used. The Morris Canal, completed in 1832 and extended in 1836, allowed for farm machinery, iron ore, lumber, and coal to be transported across New Jersey from Pennsylvania to Jersey City and Newark (Brydon, 1974; Rice, 2007). Because of their proximity to the coring sites, the industrial areas of Newark and Jersey City (Fig. 1) are also likely contributors to the deposition of atmospheric mercury from coal burning.

<u>Sample Preparation</u> Cores were subsampled for mercury at 1 cm intervals. In order to provide more material for analysis, duplicate cores of OX2A, RC2A, and TZS2A were sampled. The cores, labeled RC1A, and TZS1A were taken about 15 feet away from the main coring site for additional sampling purposes (Fig. 12). There are slight variations between these sample cores and main cores, but they are taken from approximately the same location, with mercury concentrations corresponding to the depth from the surface. Core OX1A, however, was taken on the opposite side of the oxbow, the downstream side, about 0.2 miles away from the main core site. This is because the OX1A site was the original coring site until difficulties were encountered while taking the second core. While the observed grain size and color of the two cores were nearly identical, core OX1A contained a larger amount of organics. Core OX1A was sampled to a depth of 66 cm, core RC1A was sampled to a depth of 40 cm, and core

TZS1A was sampled to a depth of 55 cm. Upon sampling, OX1A was discovered to have dried out and decreased in length by about 6 cm. Core records were used to adjust the sampling depths. Although a maximum of 0.1 grams of sample were needed for analysis, each sample was weighed out to between 0.6-0.9 grams of sample to ensure that enough sample would be available for multiple runs or accidental spills. The samples were then air dried and placed in glass sample jars.

Analysis The samples were analyzed using the Hydra IIC at the Clark Science Center at Smith College. Samples were run through the Hydra IIC mercury analyzer by Dr. Marc Anderson, who also provided assistance with sample preparation in the Smith College lab. The dried samples were ground into a fine powder either using a spice grinder or a mortar and pestle, which were wiped clean after each sample to ensure that there was no cross-contamination. Samples from 1-46 cm from core OX1A were weighed before and after analysis to determine weight percent of organic carbon from loss from combustion during the Hydra IIC analysis. However, this is not reliable for determining the exact amount of total organic carbon, as other sources of carbon (such as carbonate rocks) may also be removed during this process. While the oxbow samples contained no visible amounts of carbonates, the determined percentage of total organic carbon cannot be determined to be exact. The samples were analyzed for total organic carbon only to determine possible causes for high mercury concentrations, since mercury favors organics. Therefore, a general measurement of weight percent organic matter is sufficient. Samples from 29-42 cm of core TZS1A and 29-40 cm of core RC1A were also weighted for weight percent organics. Each sample was placed in nickel alloy sample boats and placed on sample trays that held up to 14 samples each. The Hydra IIC detects mercury using cold vapor atomic absorption and does not require any acid digestion during sample preparation. The machine was calibrated by running six empty sample boats, four acid blanks, and eight standards. The sample boat is first grabbed by the machine and inserted, allowing oxygen to flow over the sample. The sample moves to the decomposition furnace, where it is dried and combusted. The resulting gases are carried by the flowing oxygen to the catalyst furnace, removing nitrogen oxides, halogens, and sulfur oxides, and producing free mercury. Mercury and the remaining products are run through a drying tube. Mercury is then drawn into a gold amalgamation trap, which traps and concentrates elemental mercury. The trap is heated to release mercury as a carrier gas, and it is transported into the atomic absorption spectrometer (Teledyne-Leeman Labs, 2015).

HYDROLOGY

Flood Hydrographs

Discharge data for 11 selected USGS gages were used to create flood hydrographs to compare daily mean peak discharges of the Passaic River and its tributaries during major floods (Fig. 15, Table 2). The daily data and annual statistics for estimated discharge data at each of the stream gages was retrieved from the USGS National Water Information System Web Interface (USGS, 2017). Discharge data for the floods of 1902 and 1903 were taken from water supply reports (Hollister and Leighton, 1903; Leighton, 1904). Discharge data and information for floods before 1903 was obtained from historical flood reports, water supply reports, USGS peak flow data, and annual reports from the state geologist (Salisbury and Knapp, 1897; Smock, 1891; USGS, 2017; Vermeule, 1894). Gages were chosen based on the availability and length of their daily discharge records. Gages throughout the basin that were not used did not record discharge, or had limited discharge records. The Ramapo River gage at Ramapo, NY was chosen because it is located in a predominantly rural area and is used to compare streamflow rates with gages in more heavily developed areas. Flood conditions within the Great Piece Meadows are best represented by the Passaic River gage at Little Falls, which receives the combined water flow of the Rockaway, Whippany, and Pompton tributaries. Although the Passaic gage at Two Bridges is located at the edge of the Great Piece Meadows where the Pompton River joins the Passaic, and is therefore closer to the coring sites, it is a newer gage with no current discharge record. The Little Falls gage has the longest continuous USGS stream gage record, dating back to 1902, as well as flood discharge data from late 1800s reports, providing the most detailed and extensive record of streamflow in the Central Passaic Basin. Average annual discharge ranges from around 2400-199 cfs (USGS, 2017). Discharge values for the floods of 1902 and 1903 for the Little Falls gage were taken from the 1903 flood report (Hollister and Leighton, 1903; Leighton, 1904), because they give a more accurate value for the peak flood discharge. Flood discharge data for the flood of 1896 was taken from the annual report of the State Geologist for 1896 (Salisbury and Knapp, 1897) The flood of 1882 was from the 1894 annual water supply report (Vermeule, 1894). Additional flood discharge values were taken from a list of peak flows from the USGS National Water Information System website for the gage of the Passaic at Little Falls and the 1890 annual report of the state geologist (Smock, 1891; USGS, 2017). A total of 27 floods were analyzed. A list of dates of major floods were compiled from several sources since criteria for major flooding within the region were not well defined. One source, the NOAA National Weather Service Advanced Hydrological Prediction Service, listed major flooding as having a crest above 8.6-9 feet on the Passaic River at Little Falls gage (National Weather Service et al., 2016). Major floods are not solely determined by flood crests, so discharge information from NOAA and the National Weather Service was also used, which identified floods with peak discharges about or over 10,000 cfs at the Little Falls gage as major (National Weather Service, 2016). Since discharge at Little Falls was not necessarily indicative of widespread flooding within the Central Basin, lists and descriptions from the New York Division of the US Army Corps of Engineers flood risk management and restoration page and the 1987 Passaic River Mainstem Flood Protection Feasibility report and from National Water Summary 1988-89 Water-Supply Paper 2375 were used to cross reference the major floods (Paulson et al., 1991; US Army Corps of Engineers, 1987). Lastly, a modified list of maximum floods of record at the Great Falls gage in Paterson, downstream in the Lower Passaic River was used to determine major floods of the late 1800's and early 1900's (Hofman, 1955). However, major flooding within the basin did not always reflect a discharge at Little Falls of 10,000 cfs. Because it has the longest record, the gage at Little Falls was analyzed for peak discharge for major floods in the basin and for Passaic floods with a peak of ~10,000 cfs or more. Discharge data for floods before 1900 was taken from annual reports of the state geologist, except for the flood of 1869, which is from USGS (Salisbury and Knapp, 1897; Smock, 1891; USGS, 2017; Vermeule, 1894).

Fig. 15: Stream gage locations within the Passaic River Basin. Data from (NJDEP et al., 2010a; NJDEP et al., 2010b; NJOIT and OGIS, 2017; USGS, 2017)

Gage	USGS Gage Number	Drainage Area (mi²)	Location	River gradient (ft/mi) at or near stream gage (Spitz,2007)	Gradient (Spitz,2007)
Passaic River at Little Falls	1389500	762	Lat: 40°53'05" Long: 74°13'34"	10.45	.002
Whippany River Near Pine Brook	1381800	68.5	Lat: 40°50'42" Long: 74°20'50"	9.91	.0019
Passaic River near Millington	1379000	55.4	Lat: 40°40'48" Long: 74°31'44"	3.55	.00067
Passaic River near Chatham	1379500	100	Lat: 40°43'31" Long: 74°23'23"	1.04	.0002
Wanaque River at Wanaque	1387000	90.4	Lat: 41°02'39" Long: 74°17'35"	8.28	.0016
Pequannock River at Macopin Intake Dam	1382500	63.7	Lat: 41°01'06" Long: 74°24'04"	4.11	.00078
Pompton River at Pompton Plains	1388500	355	Lat: 40°58'11" Long: 74°16'55"	10.64	.00019
Rockaway River above Boonton Reservoir at Boonton	1380500	116	Lat: 40°54'06" Long: 74°24'40"	4.29	.00081
Rockaway Below Boonton Resevoir at Boonton	1381000	119	Lat: 40°54'47" Long: 74°23'36"	4.29	.00081
Whippany River at Morristown	1381500	29.4	Lat: 40°48'21" Long: 74°27'22"	10.91	.002
Ramapo River at Ramapo (NY)	1387400	86.9	Lat: 41°08'25" Long: 74°10'07"	not listed	
Passaic River Below Pompton R. at Two Bridges	1389005	740	Lat: 40°53'47" Long: 74°16'09"	0.73	.00014
Rockaway River at Pine Brook	1381200	136	Lat: 40°51'29" Long: 74°20'52"	not listed	

Table 2: USGS stream gage names and locations (Spitz, 2007; USGS, 2017)

RESULTS

CHRONOLOGY

Radiocarbon Dating

Radiocarbon results are given in years before present (B.P.), with present being 1950 with an error of +/- 30 years or +/- 32 years (Table 3). Beta Analytic determines conventional radiocarbon ages corrects the measured radiocarbon age for isotopic fractionation using δ 13C. Although radiocarbon dating is a commonly used method in dating sediment deposition, there is still a potential for error. While samples are pretreated by the lab before being processed by AMS radiocarbon dating to reduce the chance of sample contamination, there is always the possibility that the sampled organic material is not contemporary with the deposition of the surrounding sediments. Some of the radiocarbon dates are inconsistent with stratigraphy, and place younger sediments below older sediments. Seeds were included in almost all of the samples because they provide the greatest accuracy in dating contemporary sediment deposition, but the seed samples were at or below the minimum weight requirement, and may have been difficult to date. Wood is subject to the "old wood effect", in which different parts of a tree yield different radiocarbon date (BetaAnalytic, 2017). Since radiocarbon dating dates from the time organic material stops carbon exchange with the biosphere, tree rings close to the center of a tree will have a different date than the outer rings. Dead branches that remain on the tree will date differently than the rest of the tree, and their radiocarbon ages may not coincide with the dates at which they are deposited. Since seeds are deposited annually, seeds deposits can be assumed to have been dated from the year of deposition, though seeds can still be washed in from older sources, or sprout and infiltrate deeper layers. Dating multiple samples of organics within the cores provided enough consistent dates to construct a reasonable timeline for deposition within the oxbows.

Table 3: Radiocarbon sample depths and descriptions. Radiocarbon analysis and calibration by Beta Analytic and DirectAMS (Reimer and al., 2013; Talma and Vogel, 1993). Additional calibrations performed using Calib. (Reimer and al., 2013; Stuiver and Reimer, 1986-2016; Stuiver and Reimer, 1993)

Lab Number	Depth (cm)	Description	Conventional Sample Date	Calibrated Age median	Calibrated Age 2 σ	Calibrated Age 1 σ
Beta-431527	48	wood pieces, less than 5 mm, leaf fragments	5960 +/- 30 BP	6785	6730-6880	6830-6745
Beta-431528	74	small wood fragments, less than 5 mm. Few to little amounts of charcoal.	7580 +/- 30 BP	8390	8365-8415	8380-8405
Beta-445369	47	seeds	650 +/- 30 BP	655, 580, 570	555-670	565-660
Beta-409159	78.25-81	wood and charcoal, minimal seeds	2460+/- 30 BP	2680, 2640, 2610, 2600,	2360-2715	2460-2700
Beta-409160	149.25	leaves, stems, wood, minimal seeds, only 0.004 grams	4730 +/- 30 BP	5570, 5555, 5470	5325-5585	5470-5570
D-AMS 0198	176	stick, charcoal, minimal seeds	3977 +/- 32 BP	4465	4300-4526	4417-4526
Beta-405465	196.75	mostly woody material, bark, sticks, charcoal, leaves, and 0.004 g of seeds	3600 +/- 30 BP	3895	3895	3865-3965
	28-30	leaves, wood, and charcoal. Few seeds				
Beta-437085	50.5-54.5	leaf and wood fragments, sticks greater than 1 cm. dated using 10 mg of seeds	3260 +/- 30 BP	3475, 3470	3405-3565	3450-3555
Beta-431525	91.5-93.5	wood and leaves, with minimal seeds. Seeds (0.08 g)	3460 +/- 30 BP	3700	3640-3830	3650-3820
Beta-445366	105.5-108.5	sample almost all wood	3420 +/- 30 BP	3685, 3665, 3545	3590-3815	3635-3695
Beta-431526	144-147	leaves, pieces of wood, and possible roots. Seeds: 10 mg	3120 +/- 30 BP	3360	3305-3390	3270-3375
Beta-445368	153	Plentiful leaf and wood fragments, sticks 5-10 mm long, minimal seeds, less than 10 mg	3450 +/- 30 BP	3695	3635-3830	3645-3815
Beta-445367	178	wood pieces	3560 +/- 30 BP	3850	3730-3955	3835-3890

Mercury Concentrations

All three cores showed different mercury high concentration peaks at or near the tops of the cores. The causes for smaller peaks lower in the cores is unknown, but could be from influxes of sediments or groundwater from sources naturally rich in mercury. Pieces of charcoal may have been sampled and analyzed, raising mercury concentrations. Centimeter-thick pieces of charcoal were observed at the cut surface of OX1A near its concentration of 431.6 ng/g at 31 cm. A darker, less consolidated, and possibly eroded layer occurs around 36 cm, which corresponds to core RC1A's unusually high concentration of 79.6 ng/g. The correlation of charcoal and wood debris at the depth of the mercury concentration spikes in both cores suggest a forest fire as a possible reason for the high concentrations. Mercury also favors finer grained sediments, such as clay and silt (Kroenke et al., 2002). Mercury concentrations are regularly observed to decrease in the upper layers of sediment cores, likely as a result of recent emission control regulations (Lacerda and Ribeiro, 2004). Both cores OX1A and RC1A showed a decrease in concentration at the top of the core. The powdered samples were determined to be homogeneous enough that they did not require multiple runs. Increased sedimentation rates, including larger influxes of sediment due to flooding, results in the dilution of mercury soil concentrations. A sudden decrease from the average mercury concentration within the core profile could possibly indicate an increase in sediment influx.

CORE TZS2

Stratigraphy

The bottom of core TZS2 consists of angular medium-very coarse grains and pebbles (Fig. 16). The transition to dark grey (10 YR 4/1) fine-medium sand deposits starts around 168 cm (Munsell, 1994). The sand fines upwards to about 141 cm, where it transitions into very dark grey (10 YR 2/2) fine sand with some silt. The entire sand unit contains minor amounts of mica. The core fines upward from 131 cm from very fine-fine sand, to silty-sand and sandy-silt around 102 cm. From 131-102 cm, there is only a slight change in color around 107. Silt from 102- 63 cm contains high amounts of fine organics similar to peat concentrated around 73-78 cm and 64-69 cm. Wood and organics similar to peat around 74 cm were dated to 7620 +/- 30 years B.P. Two pieces of wood at 48 cm were dated to 5970 +/- 30 years B.P. A root appears around 40 cm, possibly a relic from the surface. The silt at 10-13 cm appears less consolidated and disturbed compared to the record in the rest of the core.

Grain Size

The bottom sections of Core TZS2 and Core TZS2E are not intact in stratigraphic order, since the sand was unable to be pushed out of the core barrel by the piston and had to be scooped out by hand. However, it does show a progression of upward fining sand from ~ 220-186, starting with coarse to very coarse sand and pebbles at the bottom 10 cm and fining to medium-coarse and medium sand. These sand deposits are coarser than the alluvial sand deposits in all of the cores. A preliminary sample taken by auger when first inspecting this core site was found to be medium sand (by weight percent) (Shepard, 1954). Core TZS2D continues with fining upward sand deposits, from fine-medium sand at 161 cm to fine sand with some silt at 138 cm. Core TZS2C continues with fine-very fine sand at 131 but changes to silty sand at 124 cm. It fines upward to sandy-silt and clayey-silt from 124-107 cm, but switches back to sandy-silt for 107-102 cm. The top 102 cm are made up of silt, which fluctuates in color from 10 YR 3/3 dark brown, to 1o YR 3/1 or 3/2 very dark greyish brown or very dark grey (Munsell, 1994).

Mercury

Core TZS1A shows a peak of 422.16 ng/g at a depth of 1 cm (Fig. 17B). Core TZS1A appears nearly identical to TZS2A, so no adjustments have to be made to relate mercury concentrations between the two cores. Unlike the other cores, the mercury concentration peak occurs at the top of the core, and is not followed by a decrease in mercury concentrations. This may mean that the top of the core has been eroded. The marker for coal burning is expected to occur near the top of the core, given the radiocarbon age of nearly 6,000 years at a depth of 40 cm in core TZS2A.

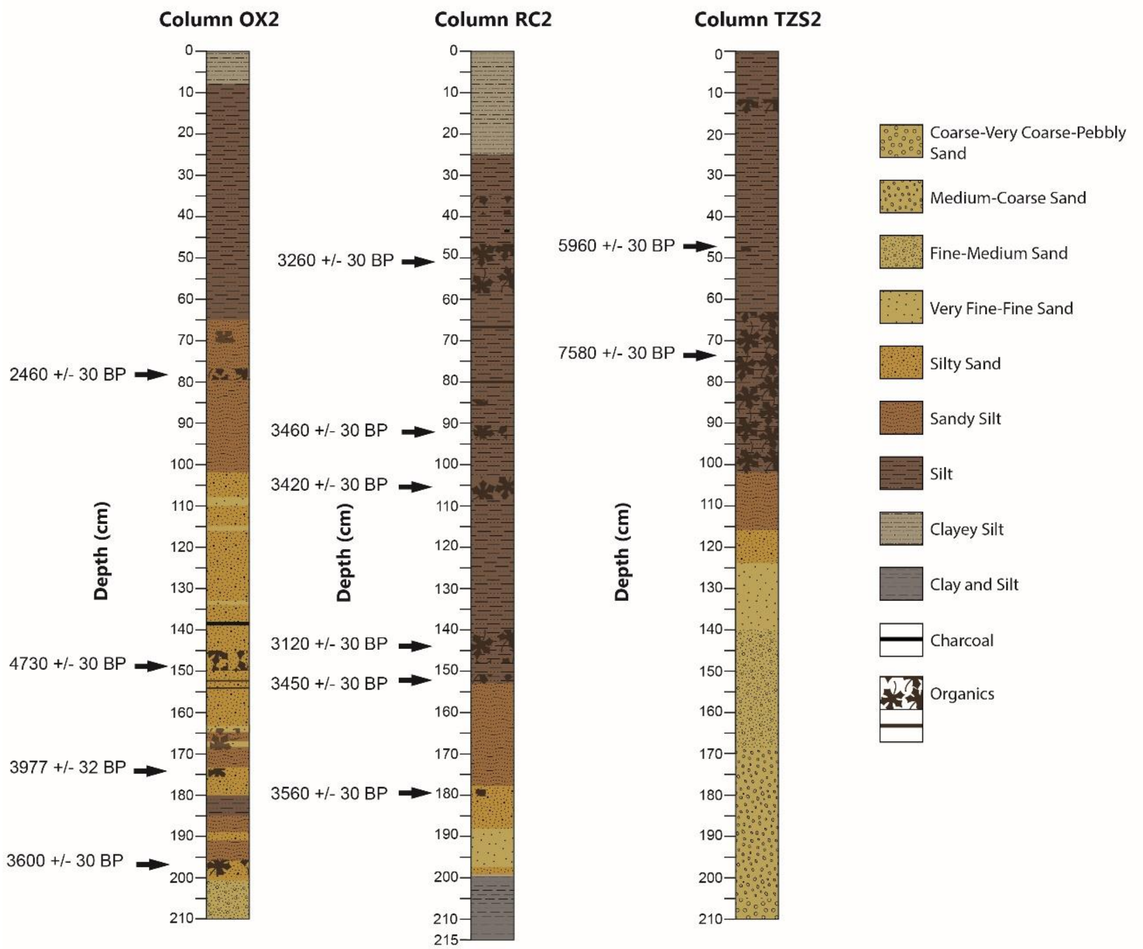


Fig. 16: Stratigraphic columns of cores Oxbow2, RC2, and TZS2 based on grain size analysis.

CORE OXBOW2

Stratigraphy

The Oxbow2 core has a general fining upward trend, beginning with fine-med sand at the bottom of the core from 210-201 cm before transitioning to silty fine-medium grained sand (Fig. 16). This boundary between the two grain sizes is gradational and difficult to determine. However, there is a sharp contrast with the darker, organic rich silty sand from 200-196.75 cm. Organics at ~ 196.75 cm were dated to 3600 +/- 30 years BP. The dark brown (7.5 YR 3/2) sandy-silt from 196.75-191.75 cm is slightly different in color than the very dark grey (10 YR 3/1) organic rich silty-sand, but contains small amounts of charcoal (Munsell, 1994). However, the silty-sand around 190 cm has no organic matter. At around 190 cm, there is a significant color change from dark yellowish brown (10 YR 3/1) to dark greyish brown (10 YR 4/2) as the sediments become more organic rich and change to sandy-silt and silt at 180.75 cm. Although the sandier sections in the lower part of the core did not contain organics, the sandy-silt to fine-med sand section from 180.25-163.15 contain fine pieces of charcoal and larger pieces of wood 4 mm-2 cm long. Although sampled twice in approximately the same location (about 171 cm), this section showed variable results in grain size as well as a significant amount of mica. While the presence of mica may have affected the results, faint bands of alternating dark gray (10 YR 4/1) and dark yellowish brown (10 YR 3/1) possibly indicate that laminations of varying grain sizes occur in this section (Munsell, 1994). Organics at 176 cm were dated to 3977 +/- 32 years B.P. Silty-sand from 163-110 cm contains fine particles of organics such as leaves and wood pieces dispersed throughout the sediment. In addition to these organics, this section of the core also contains distinct layers primarily composed of leaves and wood or charcoal. The thin layers of leaves at 153.25 and 155.25 and the layer of charcoal at 138.25 are all less than a half a centimeter thick, which the layer starting at 146 cm is about 4 cm thick. Leaves, wood, seeds from this layer were dated to 4970 +/- 30 years B.P., which is inconsistent with the material dated near the bottom of the core. This likely indicates that the organic material in this layer was washed in from another location. Thin layers of fine sand less than a centimeter thick were found at 134.25 cm and 116.25 cm. Another fine sand layer, about a centimeter thick, was observed at about 108 cm. These thin layers of fine sand and organics are important in identifying flood deposits, because they indicate a short influx of sediments of a coarser size or organic content into a long record of relatively homogenous silty-sand. The two inconsistent dates from ~150 cm and 176 cm are also indicators of flood deposits because they appear to have been washed into the oxbow from another source. A color change gradient from black (10 YR 2/1) to grey brown (10 YR 4/3) at 102-100 cm marks a shift to sandy-silt (Munsell, 1994). Sandy-silt from ~88-65 cm contain visible amounts of mica, and small particles of wood and charcoal. A layer at about 78.25-81 cm of wood, charcoal, and a small amount of 3 in depth is dated to about 2470 +/- 30 years B.P., which is consistent with the organic matter dated from the bottom section of the core. A gradual color change from very dark brown (10 YR 2/2) to very dark grayish-brown (10 YR 3/2) around 65 cm marks a shift to fine grained sediments that contain less than 20% of sand (Munsell, 1994). Two connected grayish brown (10 YR 5/5) nodules, lighter than the surrounding silt are located around 65-53 cm. These are possibly the result of bioturbation. Silt at 53-46 cm, appears mottled. Silt from 27-17 cm shows alternating dark greyish brown (10 YR 4/2) and very dark greyish brown (10 YR 2/2) bands about a centimeter in width. A slight color change from dark grayish brown (10 YR 4/2) to brown (10 YR 4/3) around 8 cm indicates the beginning of silty-clay at the top of the core (Munsell, 1994). Small amounts of charcoal are also present from 52-18 centimeters. Although there was not enough organic matter to provide a radiocarbon date, organics taken from core OX1, on the downstream side of the same oxbow, were dated to about 650 +/- 30 years B.P. at about 49 cm down. Out of the three cores, the Oxbow core has the longest transition from sand to fine grained sediments, as well as the shortest silt unit, which spanned only 53 cm. The Oxbow2 core stratigraphy was also unique in its thin layers of sand within finer grained sediments.

Grain Size

The bottom section of the complete Oxbow2 core, Ox2D progresses from in depth to fine-medium sand at 210 cm depth of the total core to silt at about 180 cm (Shepard, 1954). Grain size results from 163-180 cm varied from fine-medium sand to sandy-silt. Grain size samples also contained mica, which may have skewed results in favor of larger grains. In the second core from the bottom, OX2C, silty-sand is seen from about 110-163 cm depth, but is interspersed with thin beds less than a centimeter thick of sand (116 cm and 134 cm), organics (153 cm and 155 cm), and charcoal (138 cm). Grain size analysis of the sand layer at 134 cm showed that it contained at least 70% of sand in both samples taken but one of the samples is technically classified as a silty-sand while the other is fine sand (Shepard, 1954). Grain size analysis of the layer at 116 cm classified one of the samples as fine sand, and the other as sandy-silt. Discrepancies between the two samples taken from each of the sand layers is likely due to the small width of the layers. Some of the finer grained sediments surrounding the thin sand layers were also collected during sampling in an attempt to get a cubic centimeter of sample for grain size analysis, and would have skewed some of the grain size results to an abundance of finer sediments. Core OX2C has a thin layer of fine sand around 108 cm and silty-sand at 112 cm total depth. It then continues with sandy-silt deposits until the bottom of core OX2B. The sediments within OX2B start as sandy-silt and change around 65 cm to silt and some clayey-silt. Silt continues bottom of core OX2A (53 cm) to about 16 cm. The top of the Oxbow core (OX2A) is made up of clayey-silt.

Mercury

Core OX1A had a peak of 954 ng/g at a depth of 10 cm (Fig. 17A). Though cores OX1A and OX2A are from different locations within the same oxbow, similar features such as charcoal and sediment color occur at approximately the same depth. Samples from 1-28 cm of OX1A were run a second time because of its unusually high concentrations. The average of these two runs was used in the final analysis.

The great difference between the peak mercury concentrations of cores TZS1A and RC1A to core OX1A is unusual. The elevated concentration in OX1A must be influenced by other factors, or erosion of the other two cores removed sediments with similar mercury peaks. Core OX1A was the closest core recovered to the junction of the Pompton River and the Passaic River, so there is the possibility that its high mercury concentrations were due to contamination from cities upstream on the Pompton. Core OX1A was also shown to have high total organic carbon content which may have contributed to its high mercury concentrations due to atmospheric mercury's tendency to bond to organics matter (Appendix I, Figure 1).

A.

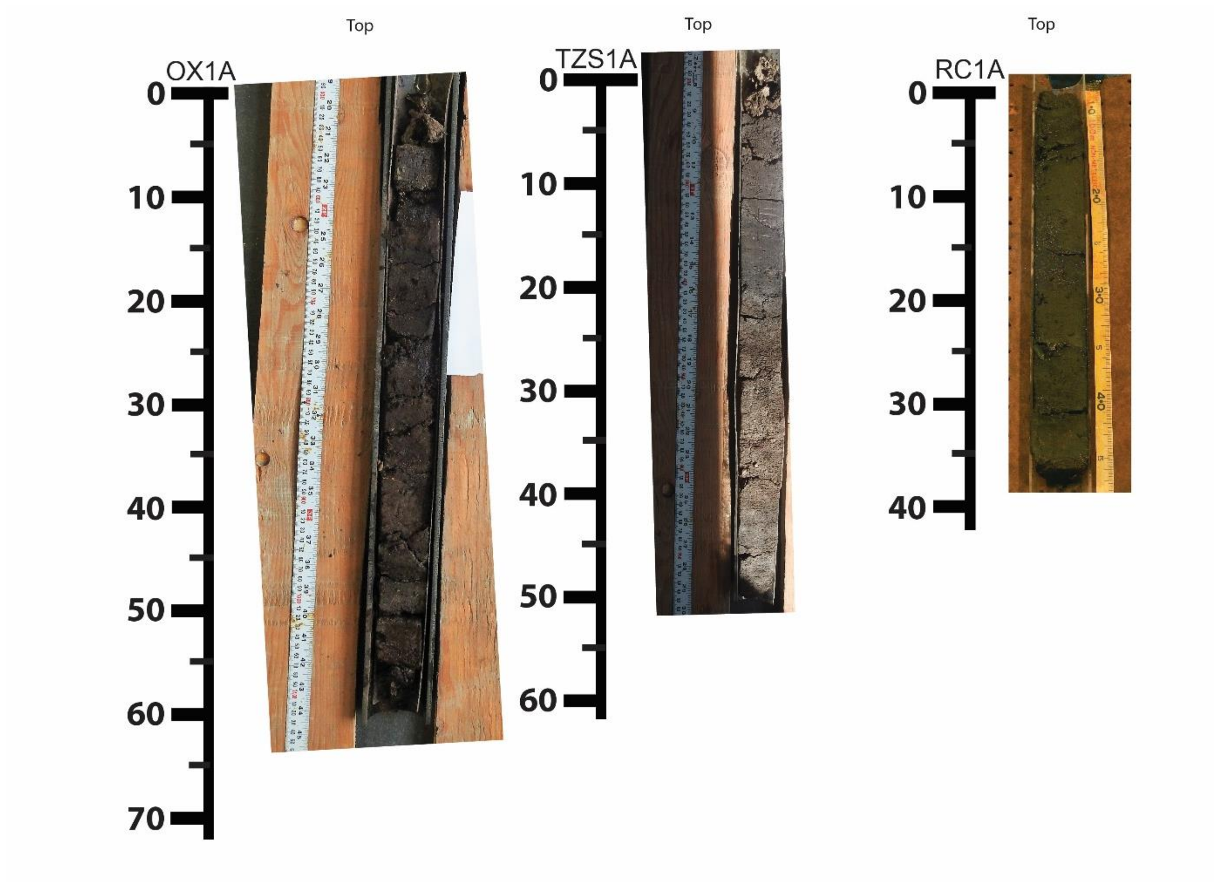


Fig. 17A: Test cores OX1A, TSZ1A, and RC1A taken for sampling.

B.

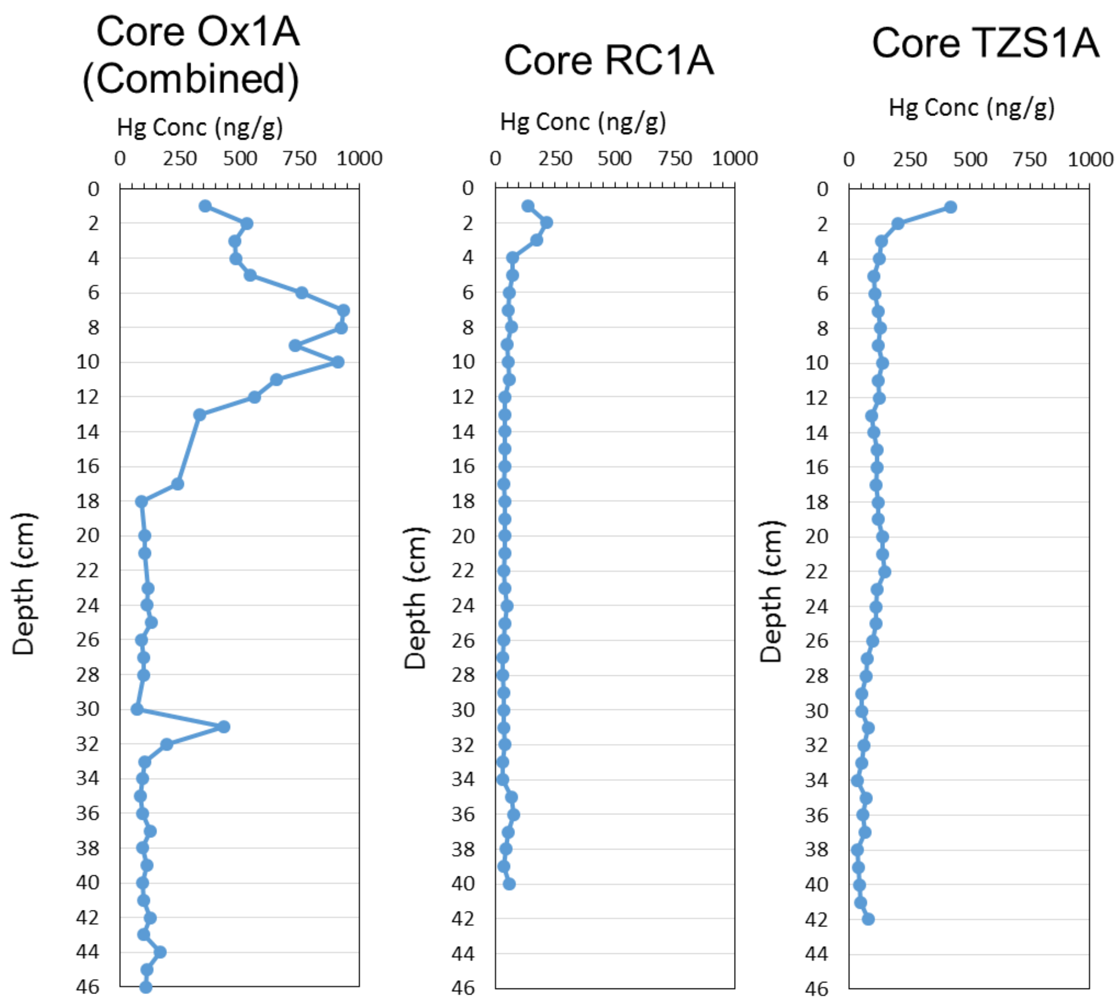


Fig. 17B: Mercury concentrations for the average of the two runs for core OX1, and for cores RC1 and TZS1

CORE RC2

Stratigraphy

Core RC is the only core to include the top of the underlying glacial lake clay deposits (Fig. 3B). Clayey silt, which is almost equal parts clay and silt, is present from the bottom of the core to 194 cm (Fig.15) (Shepard, 1954). Above the glacial lake deposits, the core begins a fining upward sequence starting with fine sand. The contact between the glacial lake silt and clay and the sand is a gradational boundary made up of sandy-silt, with a thin layer of coarse and very coarse grains. From 182 cm, the fine sand gradually transitions to silty-sand and sandy-silt before becoming silt at 151 cm. The transition from silty-sand to sandy-silt is apparent between 175 and 167 cm, where the sediment changes from dark grey (10 YR 4/1) to dark greyish brown (10 YR 3/2) and gradually becomes finer-grained (Munsell, 1994). The silty-sand and sandy-silt units both contain small amounts of mica and charcoal. Wood from 178 cm was dated to 3560 +/- 30 years B.P., and a 2 cm thick organic layer at 153 cm was dated to 3450 +/- 30 years B.P. Both of these dates are consistent with dates further up in the core. Silt from 151-140 has a gradient from very dark greyish brown (10 YR 3/2/) to very dark brown (10YR 2/2), is less consolidated than the adjacent sediments, and contains abundant woody organics (Munsell, 1994). Organics from 147- 144 were dated to 3170 +/- 30 years B.P. Silt from 132- 109.5 cm contains small amounts of charcoal and plant debris mixed with the sediments. A thick layer at 105.5-108.5 cm consisting almost entirely of wood was dated to 3420 +/- 30 years B.P. Small pieces of wood up to 2 cm in length were sampled at 92.5 cm and dated to 3460 years B.P. Thin layers of wood, about 1-2 cm thick, are located at 68.5 cm, 79 cm, and 92.5 cm. A gradual contact at about 57 cm starts a very dark brown (10 YR 2/2) organic rich layer that is and less consolidated than the surrounding sediments (Munsell, 1994). Leaves, sticks and seeds at 50.5 cm were dated to 3260 +/- 30 years B.P. A sharp contact at 47.5 cm separates the organic rich silt from silt with smaller amounts of wood, leaves and charcoal. Silt around 30 cm appears mottled and disturbed. The top 23 cm is black (10 YR 2/1) clayey-silt (Munsell, 1994). Small amounts of plant debris are scattered throughout the sediment until the top 11 cm, which contains an abundance of roots and plant material. The sediments of the RC core quickly fine upward, transitioning from fine sand to silt at a depth of 153 cm. However, it only contains a small portion of clayey silt at the top, which is similar to the Oxbow2 core, but about triple in length.

Grain Size

The bottom section of core RC (RC2C), is made up of clayey-silt from around 215 cm to 196 cm, where sandy-silt with some very coarse grains lies on top of the clayey silt. The clayey–silt was identified in the field and as the glacial clay deposits. Though categorized by the Malvern Mastersizer 3000 as containing almost equal parts of clay and silt, it is technically classified as clayey-silt (Shepard, 1954). It is possible that the Mastersizer particle analyzer under-estimated the clay fraction of the sample. This grain size sample, RC2C Sand 8, was very reactive to the hydrogen peroxide, and overflowed from its container, so it is also possible that some of the finer grain sediments were lost. A second clay sample, not treated with hydrogen peroxide, was also analyzed with the same results. This sample had the highest percentage of clay (~40%) than any other core sample, and is considered to be the upper section of the glacial lake deposits, which would have preserved coarser sediments as the lake got shallower. After the clayey silt deposits, RC2C gradually fines upwards from fine sand to silty-sand to sandy-silt before transitioning to silt at 153 cm. RC2B, which spanned from 132-34.5 cm, was comprised entirely of silt with layers of wood and organics and little amounts of charcoal throughout the core. The top core of Core RC, RC2A, continues the silt from RC2B and transitions to clayey-silt at around 25 cm. The silt appears mottled and less consolidated around 30 cm, containing both yellow (10 YR 7/6) and very dark grey (10 YR 3/1) sediments (Munsell, 1994).

Mercury

Core RC1A, located about 15 feet away from RC2A, was used for sampling for analysis of mercury concentrations. Though both cores were nearly identical, features seen in core RC1A were also observed 5 cm higher in core RC2A, so core RC2A is determined to have 5 cm of sediment eroded from the top of the core. Therefore, mercury concentrations measured in core RC1A occur 5 cm closer to the surface in RC2A . Core RC1A's peak of 216.6 ng/g at a depth of 2 cm from the surface would actually occur 4 cm above the surface in core RC2A (Fig. 17B).

Sediment classification was determined using the Shepard Sediment Classification Diagram (Shepard, 1954). Particle classification provided by analysis from the Malvern Mastersizer 3000 particle analyzer is given in percent volume. Sediment classification of samples run through the Mastersizer 3000 was determined using the values from reproducible results. Sediment classification of wet sieve samples were classified by weight percent in each sieve of the total sample. Only sample RC2C Sand 7 was analyzed by pipette analysis. For the sand samples taken from core TZS2E, only the sand was analyzed, and the sediment that passed through the number 230 sieve was labeled as a mixture of silt and clay (Fig. 16).

CAUSES OF ERROR IN GRAIN SIZE AND MERCURY ANALYSIS

Some fine grained silt and clay samples were shown by the Malvern to contain a small percentage of very coarse grains, despite inconsistencies with the majority of the sample. These very coarse grains are likely mica, which can be seen in many of the samples, because the particle analyzer assumes grains are spherical when calculating grain size (Malvern Instruments Ltd). For platy minerals such as mica or clay minerals, the laser diffraction pattern can appear wider than it does for a spherical grain of the same volume (Di Stefano et al., 2010). This can cause the particle analyzer to misinterpret irregular shaped grains as larger in volume than they are, underestimating the clay percentage of the total sample. Although no one method is perfect in determining the percent of clay sediments, the study of Goossens, 2008, listed laser diffraction analyzers such as the Malvern Mastersizer, as producing the best results (Di Stefano et al., 2010; Goossens, 2008).

Although minimal, spills and leaky test tubes caps may have resulted in some sample loss, especially of fine sediments. The only major spill observed was during pretreatment with 30% hydrogen peroxide. Sample RC2C Sand 8 was especially reactive with the hydrogen peroxide treatment, and overflowed from its test tube, reaching samples TZS2C Sand 1, Ox2A Sand 4, TZS2B Sand 1, and RC2C Sand 3. Contamination of the surrounding samples appeared to be minimal.

The samples analyzed by pipette and wets sieve analysis were subjected to possible loss of sample through spills. Since pipette analysis was not performed on the sand samples, the percent of error of the measured weights of each grain size to the total weight cannot be calculated. However, the sand samples are used to differentiate between river and floodplain deposits and the coarser grained channel deposits so minor errors in their grain size percentages does not affect the study of the Passaic River floodplain. It is also difficult to determine the percent of error of sample RC2C Sand 7 because the sample was originally intended to be analyzed by the Malvern Mastersizer 3000, and was not weighed before it was treated with 30% hydrogen peroxide solution. However, the sediment classification of this sample is consistent with the grain size analysis of adjacent samples within the core. Core OX2C was sampled twice from the same depths because six of the samples (OX2C Sand 4-9) were incorrectly marked as OX2D. While this error was quickly resolved, a second set of samples was taken to ensure reproducibility of results from OX2C samples. While there was some variability between both sets of samples from OX2C, the sample grain size results differ from their OX2D counterparts, thus confirming their source as core OX2C.

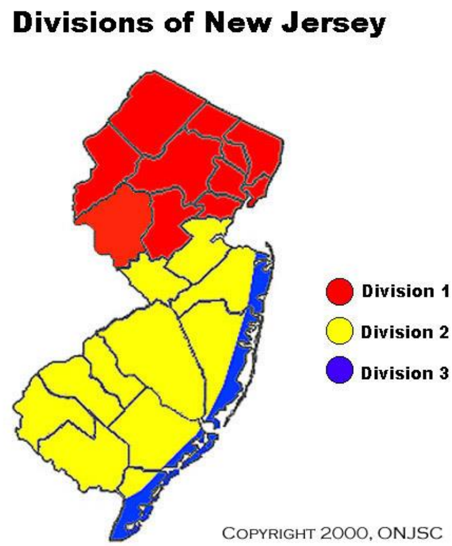
Because of constraints in time and money, only samples from core OX1 were run more than one time through the Hydra II C mercury analyzer. While the results of both runs were similar, multiple runs for each core would have ensured the reproducibility of mercury concentration results. Total organic carbon based on loss on ignition was performed for all of the OX1 mercury samples but only about half of the samples for RC1A and TZS1A. These results are sufficient for this study, which only aimed to determine the onset of widespread coal burning, but further analysis is needed to study individual mercury concentrations in depth.

HYDROLOGY DATA ANALYSIS

Precipitation

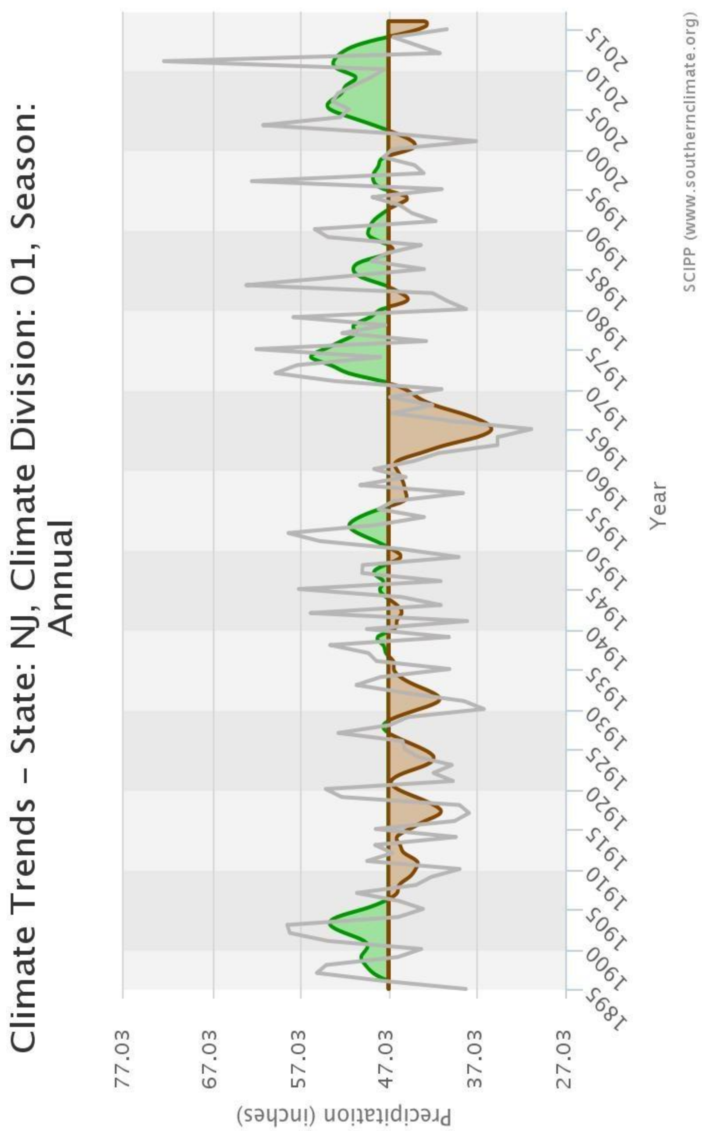
Precipitation is the primary cause for major flooding, although flooding conditions vary seasonally. Heavy storms that occur in warmer months cause major floods because they bring in large amounts of precipitation over a short amount of time. Rain during colder months may melt existing snowcover or result in increased runoff due to the frozen ground, culminating in heavy flooding (Paulson et al., 1991). Besides melting snowcover, precipitation in the months prior to flooding can raise the water levels in reservoirs and lakes, and saturate the ground, depleting any sources of floodwater storage. Precipitation data was obtained from the Office of the New Jersey State Climatologist at Rutgers University website. Data is for Climate Division 1 of New Jersey (Fig. 18), which includes Bergen, Hudson, Hunterdon, Essex, Morris, Somerset, Sussex, Passaic, Union, and Warren counties, covering 37% of New Jersey. Precipitation values are in inches of liquid equivalent precipitation, and were calculated using the NClimDiv dataset (Office of the NJ State Climatologist, 1994-2017a). The previous dataset, Drd964x, computed climate division values since 1931 by averaging the monthly data from the National Weather Service Cooperative Observer Network stations in each of NJ’s three divisions. Data before 1931 uses a different method. Statewide averages were computed by weighting the data based on the size of each division. The ClimDiv data set includes additional station data than that of its predecessor, and takes topography and network density into account, resulting in higher precision (Office of the NJ State Climatologist, 1994-2017a).

Total annual precipitation in northern New Jersey over the past 120 years has been increasing. The highest average annual precipitation on record occurred in 2011 with 72.49 inches of precipitation, 25.46 inches higher than the historical average of 47 inches (Southern Climate Impacts Planning Program, 2017). The lowest recorded annual average, 30.8 inches was 16.2 inches below the historical average, and occurred during the drought in the 1960s (Office of the NJ State Climatologist, 1994-2017a) (Fig 18). Total monthly precipitation over northern New Jersey was recorded for each of the major floods analyzed. Precipitation for the month of the flood, as well as the two months preceding, were considered to account for floods that occurred early in the month, heavy precipitation from previous months, and possible snow or ice cover that could affect the total runoff. Total monthly precipitation for northern New Jersey did not necessarily reflect precipitation conditions at Little Falls during heavy flooding (Office of the NJ State Climatologist, 1994-2017b). Therefore, monthly average precipitation for the city of Little Falls was also analyzed. The highest average annual temperature on record occurred in 2012, at 64°, 14.4 degrees higher than the historical average of 49.6 °F. The lowest average annual temperature occurred in 1904, at 55.4° (Fig. 18E) (Southern Climate Impacts Planning Program, 2017).



A

Fig 18: A: Climate divisions of New Jersey (Office of the NJ State Climatologist, 1994-2017a)



B

Fig. 18 B: Chart displays data for Division 1, Northern NJ. Long term precipitation averages are represented as a horizontal line. Annual values are plotted as a grey line. Five year moving averages are plotted against the long term temperature averages as red and blue curves. Green curves represent precipitation greater than the historical average, and brown curves represent precipitation lower than the historical average. Data from Southern Climate Impacts Planning Program website (Southern Climate Impacts Planning Program, 2017).

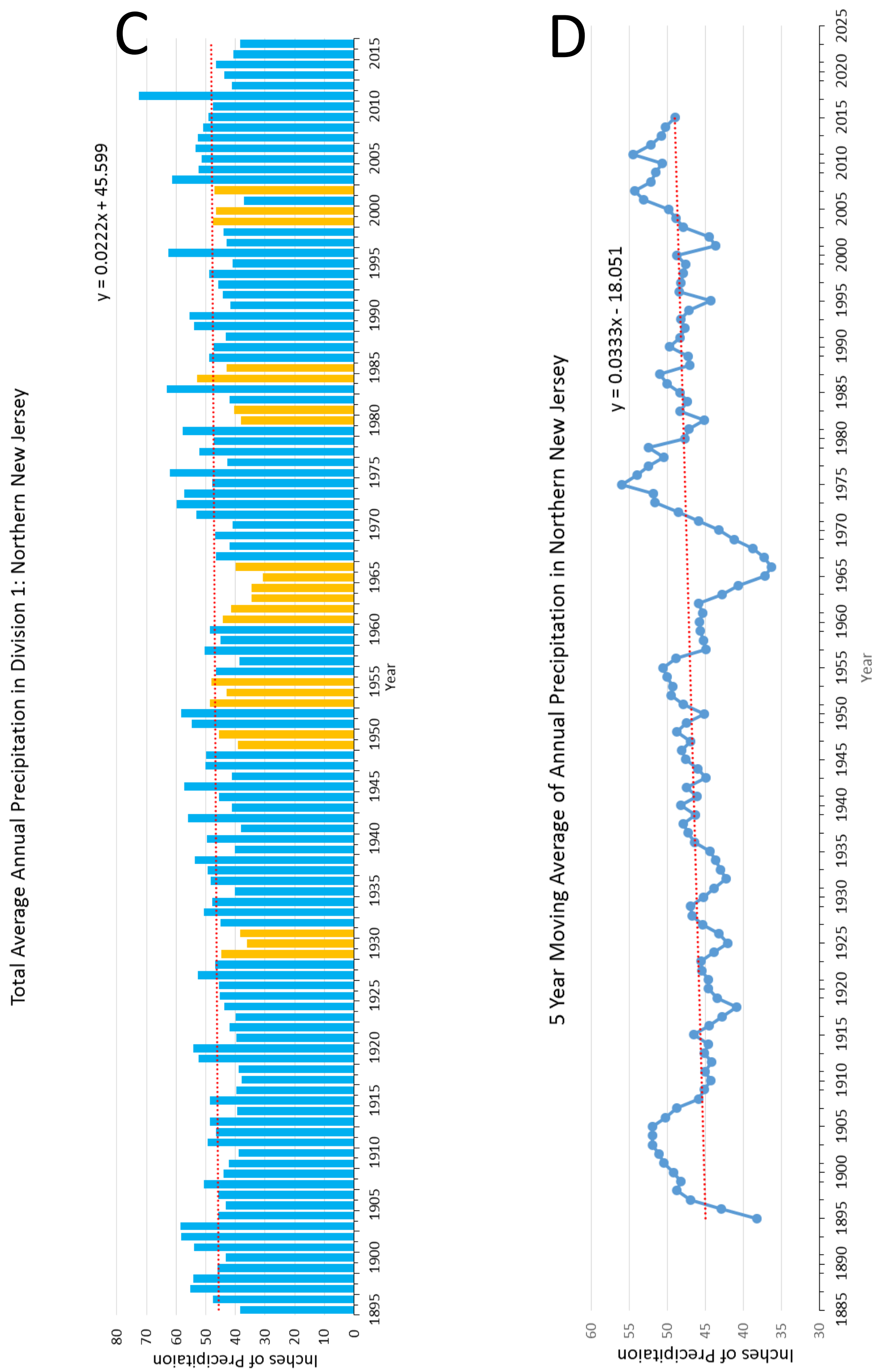
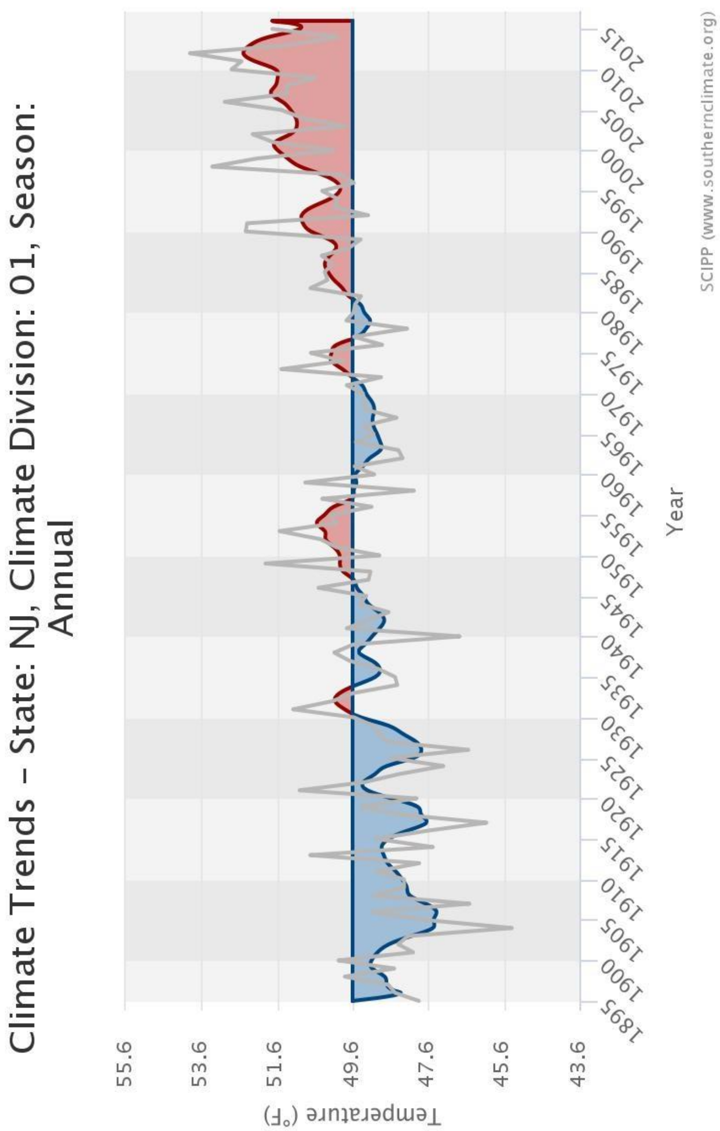


Fig. 18C: Total average annual precipitation for Division 1. Major drought years are in yellow. D: Five year moving average of annual precipitation in Division 1, starting in 1895. Data from (Office of the NJ State Climatologist, 1994-2017a).



E

Fig. 18 E: Chart displays data for Division 1, Northern NJ. Long term temperature averages are represented as a horizontal line. Annual values are plotted as a grey line. Five year moving averages are plotted against the long term temperature averages as red and blue curves. Red curves represent temperatures warmer than the historical average, and blue curves represent temperatures cooler than the historical average. Data from Southern Climate Impacts Planning Program website (Southern Climate Impacts Planning Program, 2017)

Floods

Because of its long term history of flooding problems and early industrial development, the Passaic River Basin has extensive flood and streamflow documentation since the late 1800s. Stream gages from the Upper and Central Passaic River and its tributaries provide discharge data dating back to the early 1900's. Historical flood and hydrological reports contain documentation of individual flood events, as well as discharge data from the late 1800's. Land and water use records are also available, providing context about urbanization and water diversions within the Passaic River Basin during major floods. These hydrologic records, along with climate data, and water usage records, allow for the examination of flooding patterns of the Upper and Central Passaic spanning over 100 years.

Flood Discharge

Discharge data for 11 USGS stream gages was analyzed for 27 major floods. The hydrographs show five days of average daily discharge during flooding, centered on the high peak of the Passaic River at Little Falls gage (Fig. 19)(USGS, 2017). Figure 19A shows the greatest flood to occur in over a century, and the greatest flood of record. Although data was only available for the Passaic River at Little Falls and the Pompton River at Pompton Plains gages, they record the greatest discharges on record for their gages, at 31,675 cfs and 28000 cfs respectively, though the USGS daily average peak is slightly higher for the Passaic River (Hollister and Leighton, 1903; Leighton, 1904; USGS, 2017). Precipitation data for the town of Little Falls is over 4 inches higher than the regional precipitation values (Office of the NJ State Climatologist, 1994-2017a; Office of the NJ State Climatologist, 1994-2017b). It is the only other flood besides the flood of July 1945 that shows peak discharges for both the Pompton Plains and Little Falls gages on the same day (USGS, 2017).

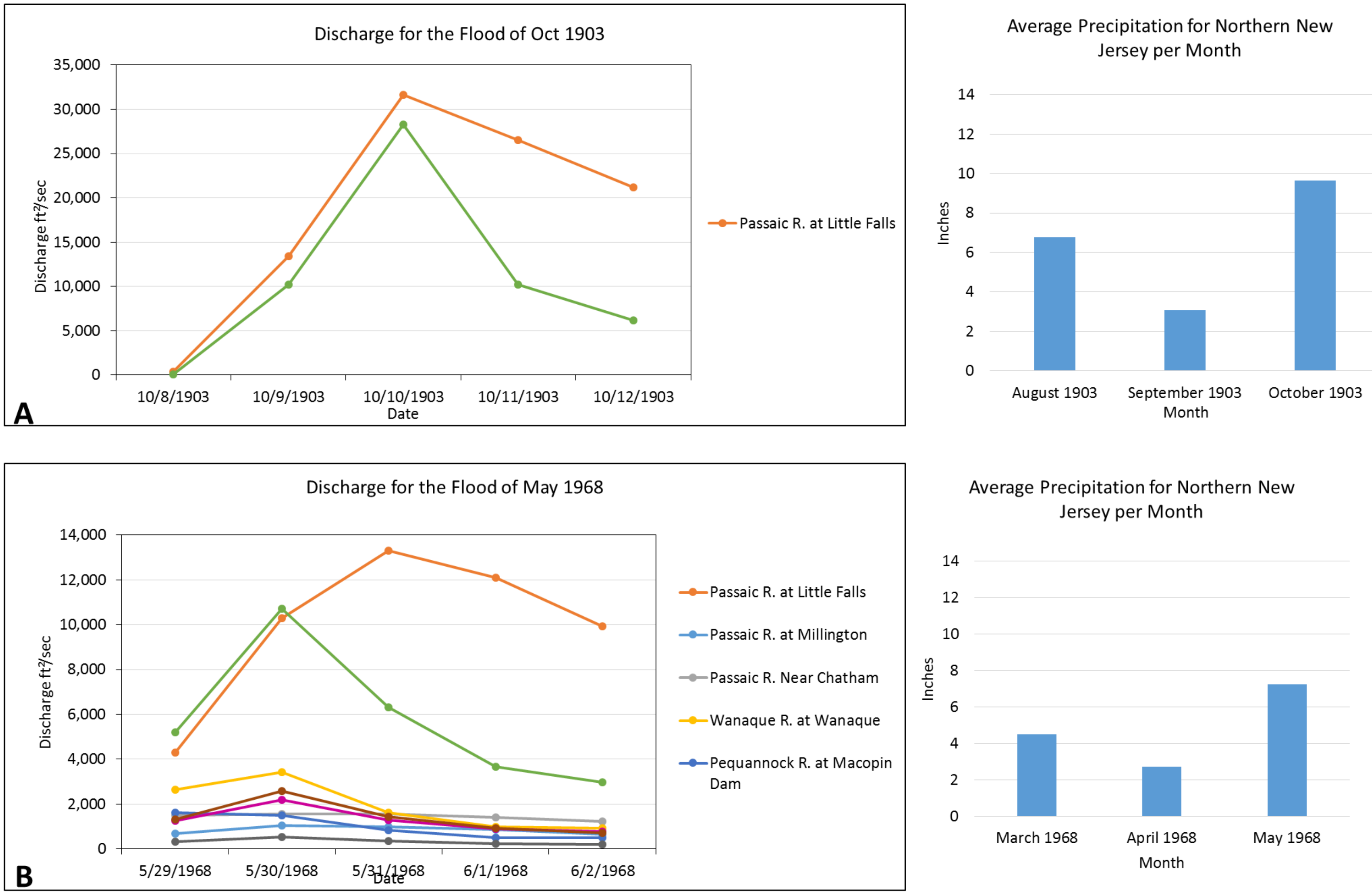
The flood of 1968 (Fig. 19B), shows the average daily discharges of the tributaries of the Passaic and those of gages upstream peaking one day before the Passaic River at Little Falls. Regional precipitation was 7.2 inches, and 8.9 inches at Little Falls for the month of May (Office of the NJ State Climatologist, 1994-2017a; Office of the NJ State Climatologist, 1994-2017b). The Passaic River at Millington, the gage furthest upstream of the Passaic, showed more variation in discharge than the Passaic near Chatham, which had little change. Both gages above and below the Boonton Reservoir peaked on the same day; the gage below the reservoir having the highest discharge (USGS, 2017). The Wanaque River at Wanaque gage had the highest discharge of the tributaries besides the Pompton. Both reservoirs were at a high capacity during the month of the flood.

The flood of April 1984 (Fig. 19C) was the first of two major floods in 1984. Regional precipitation was at 5.83 inches during April, while precipitation at Little Falls was slightly higher at 7.57 inches, with similar values for the preceding month (Office of the NJ State Climatologist, 1994-2017a; Office of the NJ State Climatologist, 1994-2017b). The Pompton gage was had a significantly high average daily peak discharge near 20,000 cfs, which the Little Falls gage had a lower (but still high) discharge peak a day later at 18,000 cfs (USGS, 2017). The Ramapo River and Whippany River at Morristown gage peaked the earliest. The Ramapo gage had the highest peak of the remaining gages, while the Wanaque River and Rockaway River gages were notably high as well. The Upper Passaic gages showed barely any change. Both the Boonton and Wanaque reservoirs were at high capacity during the flood.

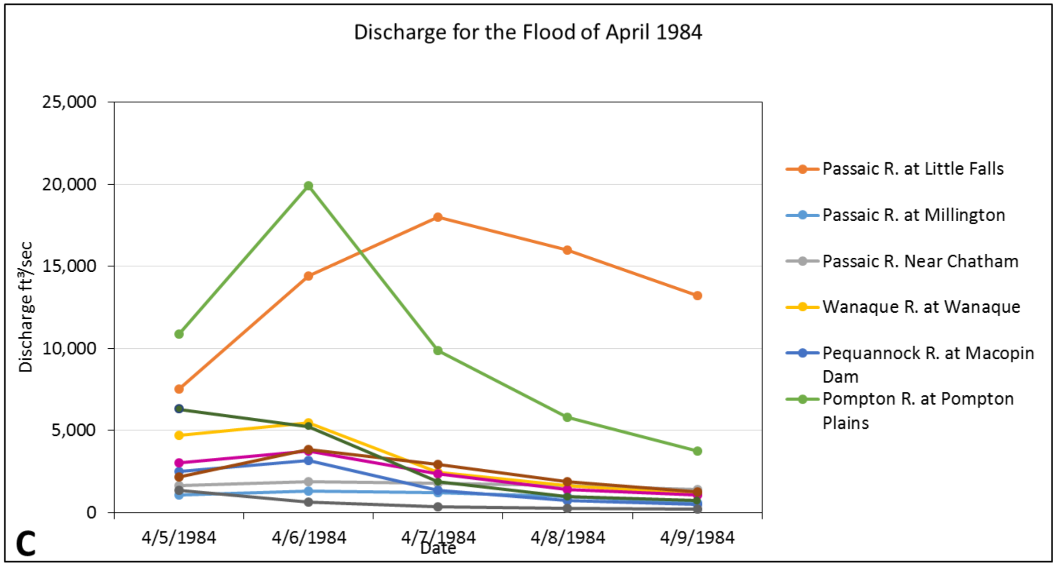
Precipitation during the month of the flood of January 1979 (Fig. 19D) was high at over 10 inches, with a similar amount at Little Falls (Office of the NJ State Climatologist, 1994-2017a; Office of the NJ State Climatologist, 1994-2017b). Precipitation during the preceding months was less than half that amount. All of the tributary and up river gages except Wanaque River at Wanaque peak two days before the Little Falls gage. The Pompton and Rockaway River gages rapidly decline in discharge after peaking. The Wanaque reservoir was not at peak capacity (near 30,000 MG) at 27,408 MG, but it was not at a greatly lowered capacity either. While close at 6318 MG, the Boonton Reservoir was also not at peak storage capacity (around 7,700 MG) (NJGS, 2013).

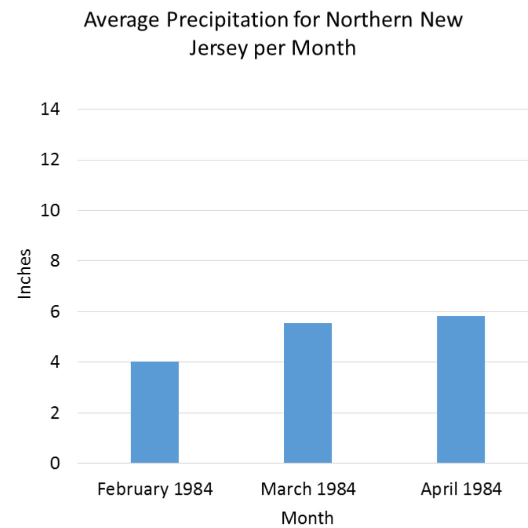
During the month of September in 1999 (Fig. 19E), regional precipitation was at 11.21 inches (Office of the NJ State Climatologist, 1994-2017a). And precipitation during September at Little Falls reached almost 18 inches. However, peak discharge at Little Falls remains under 12,000 cfs. While high amounts of precipitation did not greatly affect the major flood discharge at Little Falls, it does produce high discharges at the Pompton Plains gage, the Rockaway River above Boonton Reservoir gage, and the Ramapo River at Ramapo gage, which peak two days before the gage at Little Falls (USGS, 2017). The earliest gage to peak is at Whippany at Morristown. Both the gages below the Wanaque and Boonton Reservoirs show little change in daily average discharge.

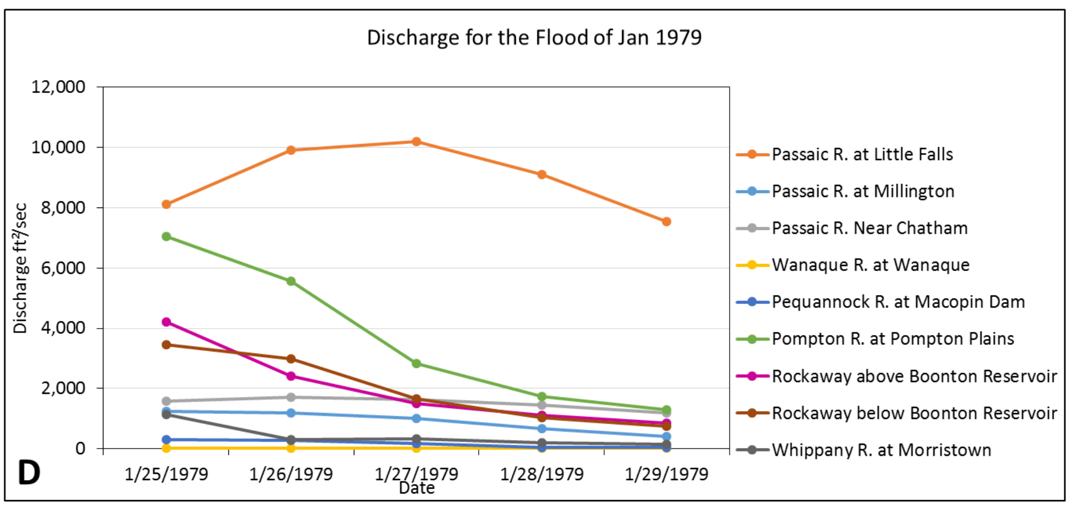
The first flood resulting from Hurricane Irene (Fig. 19F) occurred after the month with the highest precipitation of the analyzed floods. This resulted in the highest discharge peak at Little Falls after 1950, and the highest peak at the Pompton River at Pompton Plains gage after the flood of 1903 (USGS, 2017). The Ramapo River, the Rockaway River above Boonton, and the Whippany River at Morristown gages all peak two days before the Little Falls gage. The Rockaway River below Boonton Reservoir gage peaked one day before Little Falls. Though Boonton Reservoir data does not continue into 2011, the Wanaque Reservoir was shown to be at a high capacity (NJGS, 2013).

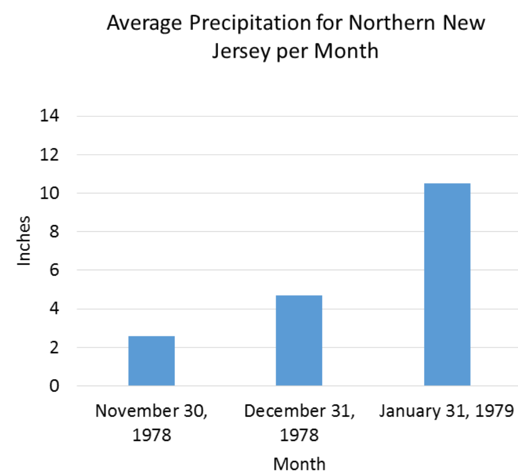


Discharge for the Flood of Oct 1903
Discharge ft²/sec
35,000
30,000
25,000
20,000
15,000
10,000
5,000
0
10/8/1903
10/9/1903
10/10/1903
10/11/1903
10/12/1903
Date
Passaic R. at Little Falls
A
Average Precipitation for Northern New Jersey per Month
Inches
14
12
10
8
6
4
2
0
August 1903
September 1903
October 1903
Month
Discharge for the Flood of May 1968
Discharge ft²/sec
14,000
12,000
10,000
8,000
6,000
4,000
2,000
0
5/29/1968
5/30/1968
5/31/1968
6/1/1968
6/2/1968
Date
Passaic R. at Little Falls
Passaic R. at Millington
Passaic R. Near Chatham
Wanaque R. at Wanaque
Pequannock R. at Macopin Dam
B
Average Precipitation for Northern New Jersey per Month
Inches
14
12
10
8
6
4
2
0
March 1968
April 1968
May 1968
Month


Discharge for the Flood of April 1984
Discharge ft³/sec
25,000
20,000
15,000
10,000
5,000
0
4/5/1984
4/6/1984
4/7/1984
4/8/1984
4/9/1984
Date
Passaic R. at Little Falls
Passaic R. at Millington
Passaic R. Near Chatham
Wanaque R. at Wanaque
Pequannock R. at Macopin Dam
Pompton R. at Pompton Plains
C


Average Precipitation for Northern New Jersey per Month
Inches
14
12
10
8
6
4
2
0
February 1984
March 1984
April 1984
Month


Discharge for the Flood of Jan 1979
Discharge ft²/sec
12,000
10,000
8,000
6,000
4,000
2,000
0
1/25/1979
1/26/1979
1/27/1979
1/28/1979
1/29/1979
Date
Passaic R. at Little Falls
Passaic R. at Millington
Passaic R. Near Chatham
Wanaque R. at Wanaque
Pequannock R. at Macopin Dam
Pompton R. at Pompton Plains
Rockaway above Boonton Reservoir
Rockaway below Boonton Reservoir
Whippany R. at Morristown
D


Average Precipitation for Northern New Jersey per Month
Inches
14
12
10
8
6
4
2
0
November 30, 1978
December 31, 1978
January 31, 1979
Month

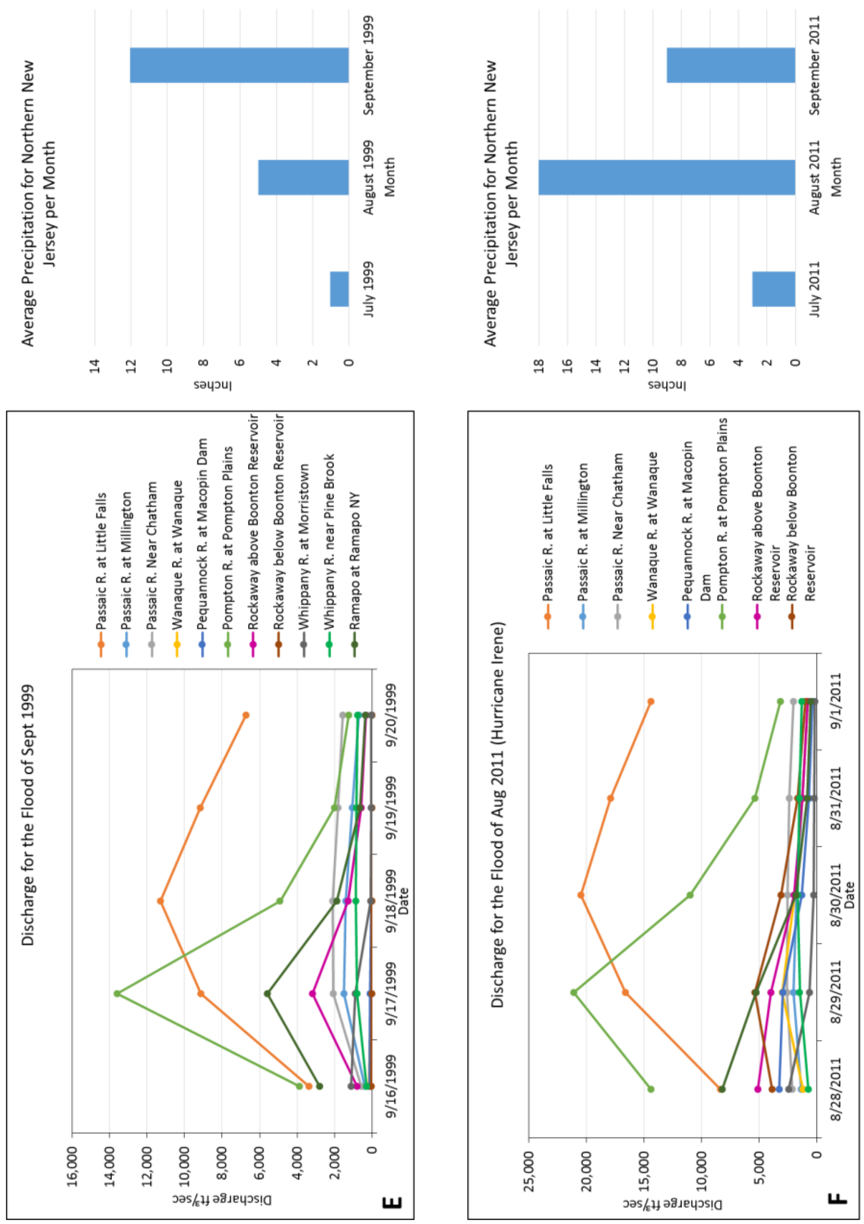


Fig. 19 A-F: Peak recorded discharge values at each gage and average monthly precipitation for the months during and preceding six major floods. Data from (Office of the NJ State Climatologist, 1994-2017a; USGS, 2017; Leighton, 1904; Hollister and Leighton, 1903)

Reservoirs and Water Usage

Reservoirs and water diversions within the Passaic River Basin also factor into regional flooding (Fig. 20) (Table 4). While they may not prevent flooding, major reservoirs can influence the severity of flooding depending on how full they are around the time of the flood. The Passaic Basin has ten major reservoirs, the largest being the Wanaque Reservoir (Table 5)(USGS, 2005). Most of them have existed since the late 1800s or the early 1900s, with only two reservoirs completed after the 1920s (Table 5). Data compiled by the New Jersey Geological Survey includes storage data from the reservoirs recorded by the major water purveyors supplying the population of northeastern New Jersey (Table 4)(Figs. 21 and 22) and monthly data of water diversions either from the rivers to the water supply companies or reservoirs, or from reservoirs for water supply distribution (Fig. 24) (NJGS, 2013). A more intensive study is needed to determine the individual effects that specific diversions and reservoir storage amounts have on flooding within the Passaic River Basin, general trends can be determined and compared to average annual and major flood discharge.

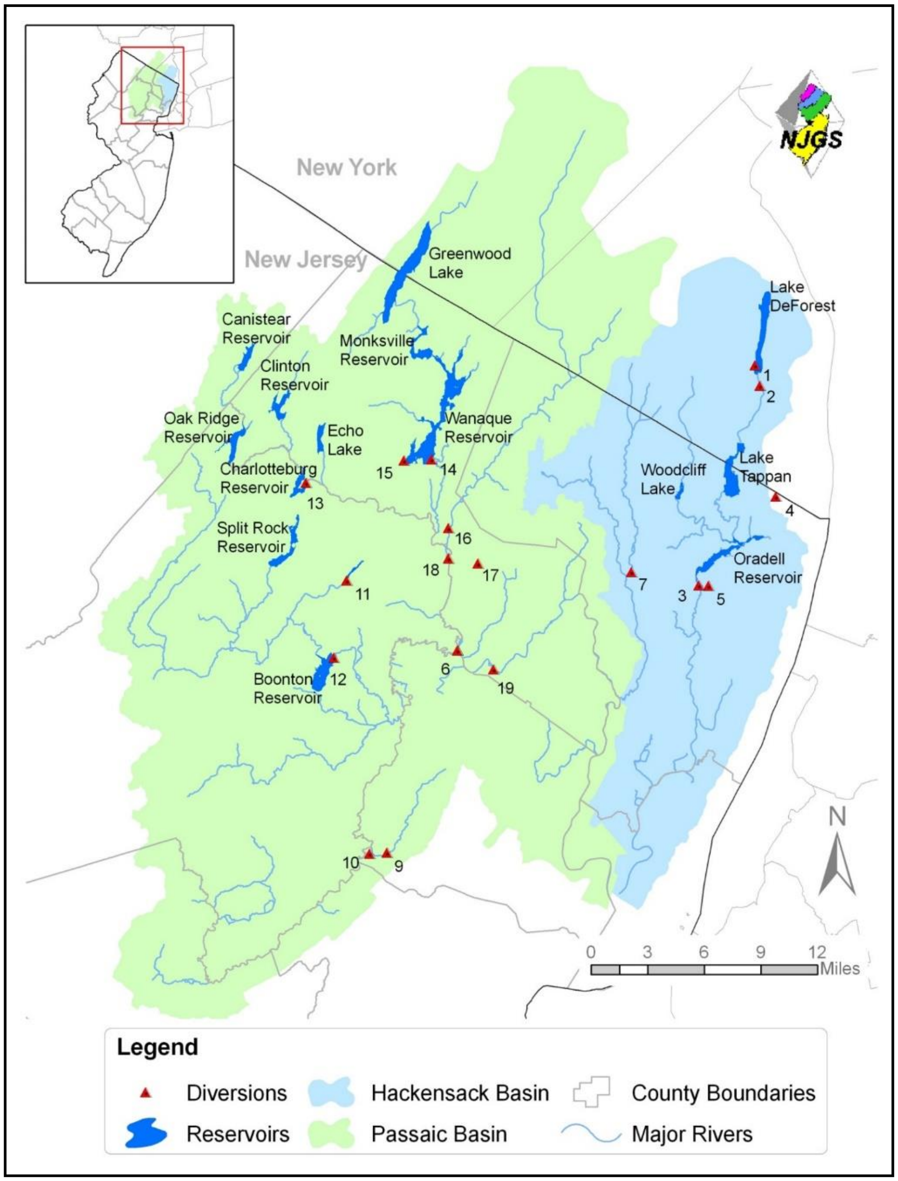


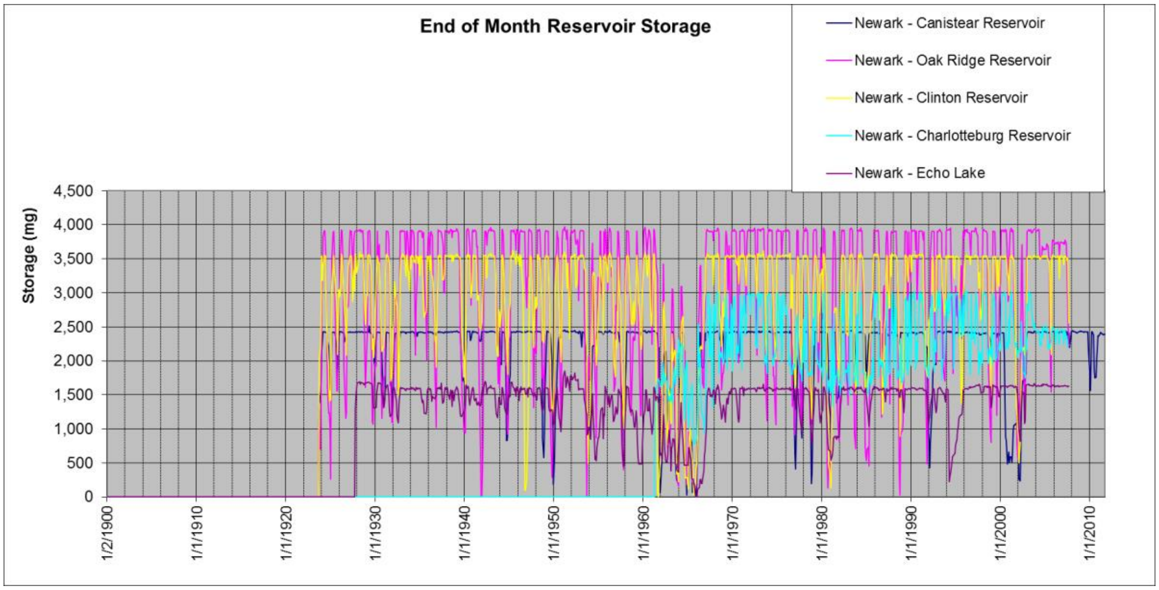
Fig. 20: Reservoirs and water usage diversions within the Passaic River Basin. Passaic Basin diversions highlighted in green. Data from New Jersey Geological Survey (NJGS) Digital Geodata Series (NJGS, 2013)

Map Label	Diversion Name	Diversion Owner	USGS_ID
1	De Forest Lake	United Water NY	01376699
2	Hackensack River	West Nyack NY	01376810
3	Oradell Reservoir	United Water NJ	01378478
4	Sparkill Creek [1]		01376272
5	Hirshfield Brook		01378521
6	Pompton River		01388981
7	Saddle River		01390520
8	Wells to Surface Supply		multiple locations
16	Ramapo River		01387990
9	NJAWC Canoe Brook	NJ American Water Company	01379530
10	NJAWC Passaic River		01379510
11	Stony Brook tributary Diversion at Taylortown	Town of Boonton	01380280
12	Boonton Reservoir	Jersey City	01380800
13	Charlotteburg Reservoir	Newark	01382370
6	Pompton River to Wanaque Reservoir	North Jersey District Water Supply Commission	01388980
14	Wanaque Reservoir		01386980
15	Post Brook to Wanaque Reservoir		01387020
16	Ramapo River to Wanaque Reservoir		01387990
6	Pompton River to PVWC at Little Falls	Passaic Valley Water Comission	01388982
17	Point View Reservoir to Little Falls		01387959
18	Pompton River to Point View Resevoir		01388490
19	Passaic River to PVWC at Little Falls		01389490

Table 4: Water Diversions of the Passaic River Basin (NJGS,2013)

Reservoir	USGS Number	Date Completed	Capacity (gal)
Splitrock Reservoir	1379990	1925	3,310,000,000
Boonton Reservoir	1380900	1904	7,620,000,000
Canistear Reservoir	1382100	1896	2,407,000,000
Oak Ridge Reservoir	1382100	1892	3,895,000,000
Clinton Reservoir	1382300	1892	3,518,000,000
Charlotteburg Reservoir	1382380	1961	2,964,000,000
Echo Lake	1382400	1925	1,583,000,000 with additional storage of 180,000,000 with flashboards
Greenwood Lake	138300	1837, recompleted 1928	6,860,000,000 with 7,140,000,000 dead storage
Monksville Reservoir	1384002	1988	7,000,000,000
Wanaque Reservoir	1386990	1927-1928	29,630,000,000

Table 5: Data obtained from USGS Reservoirs in Passaic River Basin, USGS Publications Warehouse (USGS, 2005)



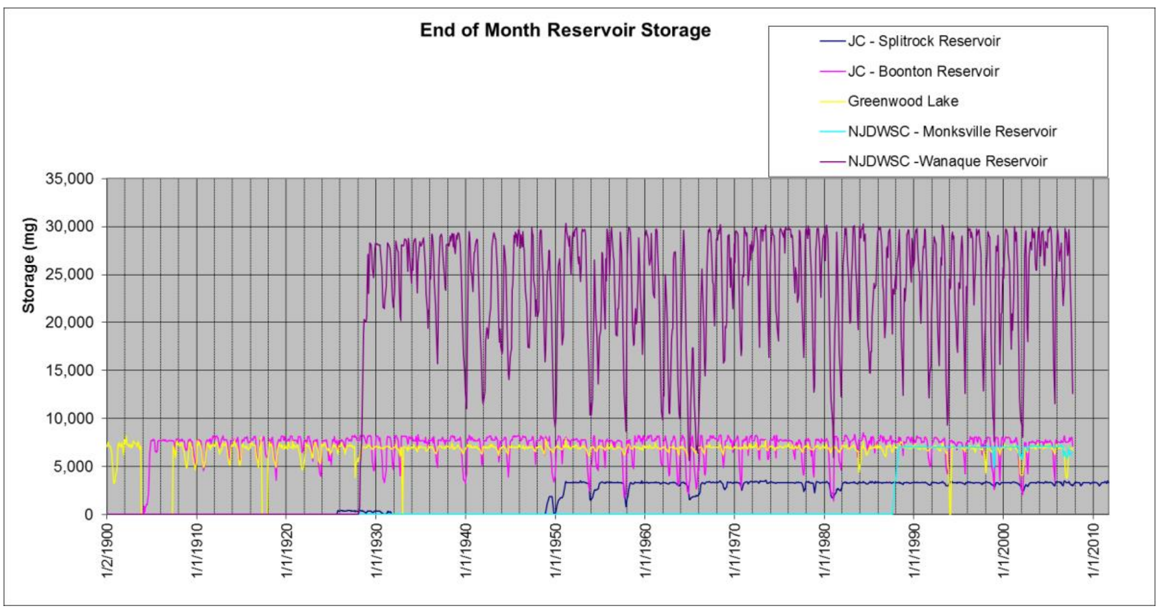


Fig. 21: End of month storage statistics for reservoirs run by Jersey City and North Jersey District Water Supply Commission. Data from New Jersey Geological Survey (NJGS) Digital Geodata Series (DGS) (NJGS, 2013).

<u>Runoff Ratio</u>

To determine how much of precipitation made it to the rivers as runoff, annual runoff ratio estimates were calculated for the drainage areas of each of the stream gages. The runoff ratio is a dimensionless proportion of the precipitation on the drainage area that has not infiltrated into underlying aquifers or removed by evapotranspiration to the discharge of the river measured at the gaging stations. The annual runoff ratio for each stream gage was calculated by dividing the average mean discharge for the entire year (cfs/year) at each stream gage by the average annual precipitation (ft/year) for northern New Jersey over the drainage area (ft^2) of each gage (Office of the NJ State Climatologist, 1994-2017a; USGS, 2017). Runoff ratios are mainly affected by surficial sediment permeability, topography, and urban development (Minnesota Pollution Control Agency, 2015). Possible factors affecting the runoff ratio within the Passaic River basin are increases over the past century in impervious surfaces such as parking lots and roads, the construction of dams and reservoirs along the river, and deforestation as a result of urban development. The differences in topography between the Highlands and Central Basin river watersheds and their surficial geology also factor into the runoff ratios of each river. While most dated back to the early 1900s, the length of the annual discharge record varied among gages. However, several trends of runoff ratios values were apparent in the recorded stream gage values.

DISCUSSION

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STRATIGRAPHY AND SEDIMENTOLOGY

Sedimentary Record of Oxbow Cores

Much of the geomophological history of the postglacial Central Passaic River floodplain can be told by the deposits in the three oxbows. The oxbows themselves are an interesting feature of the floodplain, since the shallow gradient of the river and the hundreds of acres of absorbent wetlands makes avulsion an unlikely event. The area in which the oxbows are located is labeled as alluvium on the surficial map of the Pompton Quadrangle; a layer ranging from silty-clay to gravel that rests on top of glacial lake deposits and can be up to 20 feet (6 m) thick (Stanford, 2007b). However, this study aims to define this alluvium layer on a smaller scale, using oxbow cores about 2 meters in length. Though they vary in flood deposits and timing of their meander cutoffs, all three oxbow cores show a general fining upward sequence. Postglacial sediments in each core begin with channel sand, transitioning to silty-sand and sandy-silt before becoming completely silt. The Oxbow2 and RC2 cores are interspersed with organic rich sections and fine sand and fine organic laminae deposits and are capped by a silty-clay layer. Glacial lake deposits of silt and clay are only present at the bottom of core RC2, but their existence was confirmed at the bottom of the sand layer of oxbow TZS by augering. Some degree of compaction is expected, although thought to be minimal due to the type of sediment within the cores. Since compaction cannot be accurately measured, so sedimentation rates can only be estimated. Radiocarbon dates are the most reliable in establishing a timeline of deposition.

Chronology

Radiocarbon Dates Though the cores are interbedded with thin layers of flood deposits, no obvious markers exist that would correlate flood deposits among the three cores, or link them to known flooding events. Therefore, multiple samples of organic matter were taken from each of the oxbow cores for radiocarbon dating in order to establish a timeline of events for the three oxbows. However, the resulting dates are not all consistent with the age of deposition, so errors had to be accounted for during the interpretation in the timeline of oxbow sedimentation.

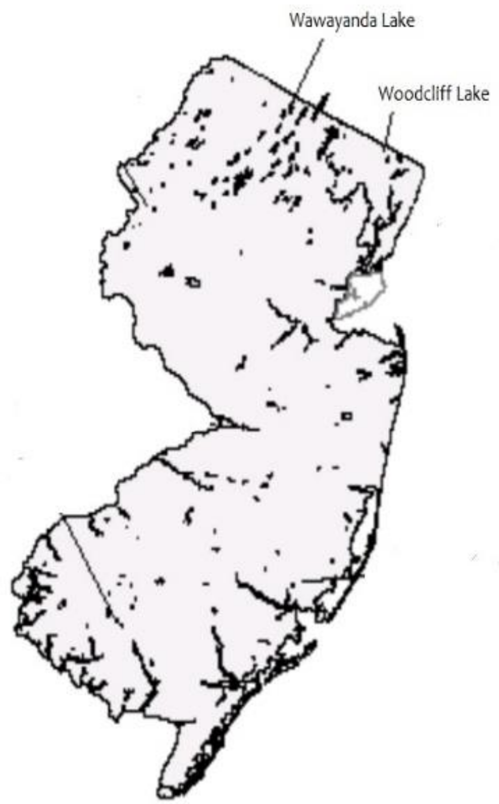
Samples of woody materials and leaves deemed to be inconsistent with the dating sequence of the core are likely organic material washed in from another location upstream by high, fast moving waters, and can be considered indicators of periods of intensive flooding. Although core TZS2 only yielded two radiocarbon dates, its organic deposits of wood, or silt containing a large amount of fine organics; similar to peat, but with less organic content. These peat-like organics are more likely to have remained in place in the oxbow rather than have been washed in, and can therefore be expected to yield accurate dates.

One issue with a sample in core RC2 gave organic matter, including seeds, from 147- 144 a date of 3120 +/- 30 years B.P., which is younger than the organic matter dated above the sample within the sedimentary record. This may have been due to a radiocarbon dating error or sediment reworking. Possible remnants of roots were also found within this organic layer, which may have infiltrated from a plant that grew years after this layer was deposited, and therefore yielded a younger date. It is also possible that younger sediments closer to the surface were accidentally collected during the descent of the core barrel.

Because of a lack of a sufficient amount of large organic deposits, no radiocarbon dates were determined for the top 40 cm of each core. However, organics were taken from core OX1, a core located within the same oxbow as the Oxbow2 core. These were dated to about 650 +/- 30 years B.P. at about 49 cm depth. Due to its high organic content, core OX1A was measured to have had around 50% compaction, but this may not be a realistic value because of issues measuring compaction with the Livingston corer. However, mercury concentrations indicated that the start of widespread coal burning, around 1835, is recorded in the core at around 18 cm. This is at a similar depth to the increase of mercury concentrations in core RC, making an age of 650 +/- 30 years B.P at a reasonable date for sediments at about 49 cm depth.

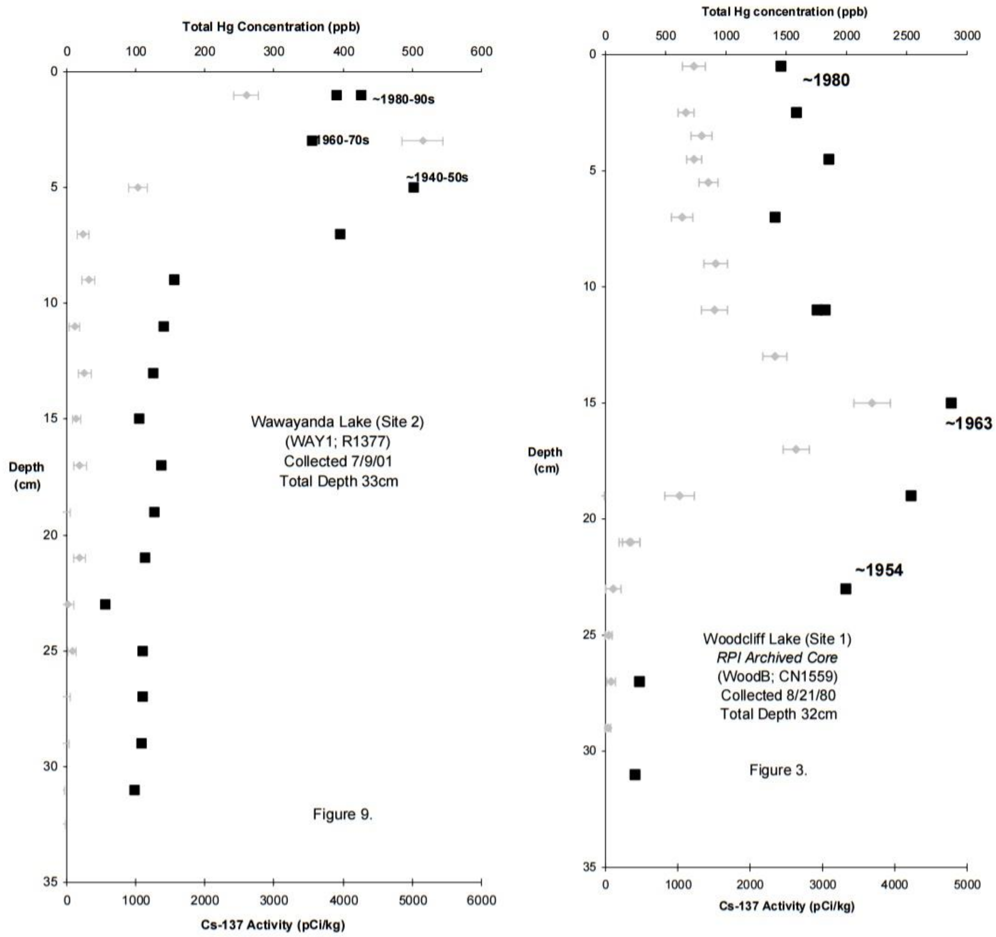
<u>Mercury Past Studies</u> Mercury concentration measurements were used to date the upper sections of each core. While it is still difficult to accurately label the start of coal burning industries (~1835) within the sedimentary record of the Great Piece Meadows, mercury concentration measurements allow for a good estimate. Because the cores could not be dated using Pb-210 because they lacked the steady sedimentation of a lake environment, data from similar mercury studies in the surrounding regions and were used to help interpret where the beginning of coal burning (~1835) is reflected within the core. For Kroenke et al’s (2002) final report to the NJDEP which chronicled atmospheric mercury deposition at six lakes sites throughout New Jersey. Five of the six sites had

mercury fluxes comparable to those in the Great Lakes, with atmospheric deposition at about 150 µg/m²/yr (Kroenke et al., 2002). The two lake sites that fall within or near the Passaic River drainage basin are the Wawayanda Lake and Woodcliff Lake site (Fig. 22).



A.

Fig. 22A: Wawayanda and Woodcliff Lake cores locations (Kroenke et al., 2002).



B.

Fig. 22B: Wawayanda and Woodcliff Lake cores mercury concentrations. The background error bars is Cs-137 activity (pCl/kg) (Kroenke et al., 2002)

Woodcliff Lake, located within the Hackensack River Basin, was dammed in 1906, so it does not record the onset of industrial coal burning. Core WoodB is 32 cm long, and displays a peak in Cs-137 activity at the 14-16 cm segment, which correlates to the years 1963-1964. An average net sedimentation rate of 1 cm/yr was calculated for the period of 1963-1980, and 0.9 cm/yr between 1954-1980 (Kroenke et al., 2002). The cesium profile is smooth, indicating continuous sedimentation. Mercury concentrations show a peak of ~2870 ng/g at a depth of 15 cm, which correlates to a date of ~1963. Peak Hg flux also occurs at this depth. Peak values of mercury accumulation within the Northern hemisphere are commonly attributed to the 1960s and 1970s (Lacerda and Ribeiro, 2004). Sandy sections at the bottom of the core have concentrations of ~245 ng/g. WoodD, another Woodcliff Lake core taken from the deepest section of the lake (~25 ft), and is believed to be an extension of WoodB. WoodD is a 32 cm long core with Hg concentrations ranging from 293-359 ng/g, and has a calculated sedimentation rate of 1.6 cm/yr. The Woodcliff lake site has abnormally high mercury concentrations. The cause of this is unknown, but this may be caused by its proximity to New York City, which provides atmospheric and solid waste contamination (Kroenke et al., 2002).

Wawayanda Lake was dammed in 1846 to create a large lake from two smaller lakes. WAY1, a 33 cm core, was collected from the deepest part of the lake (~61 ft) (Kroenke et al., 2002). Measured total mercury concentrations ranged from 55-502 ng/g, peaking around 5 cm, which correlates to the 1940-1950s. Cs-137 peaks at a depth of 2-4 cm, indicating very low sedimentation rates at the top of the core. A calculated constant depth sedimentation rate of ~0.1 cm/yr was calculated, the lowest flux in the study.

Mercury analyses from previous studies are also comparable to the results from the Great Piece Meadow cores. Concentrations in more forested areas of New York, such as the Adirondacks, had total mercury concentrations ranging from 80-500 ng/g (Lorey and Driscoll, 1999), while urban areas, such as in Central Park, Manhattan, had high concentrations of 1000-2000 ng/g (Kroenke et al., 2002). The National Water Quality Assessment analyzed three sediment cores for mercury from northeastern New Jersey as part of a program conducted in 1997 (Bond et al., 2001). Results from these cores showed total Hg concentrations that ranged from 230-5130 ng/g. The highest of these concentration came from Orange Reservoir, within an urbanized watershed which lies just outside of the Passaic River drainage basin. Clyde Potts Reservoir, which lies inside the drainage basin, upstream of the Great Piece Meadows, is located within a more forested area and has Hg concentrations ranging from 260-380 ng/g (Van Metre and Callender, 1997). Packanack Lake, although located in a more urbanized area, lies a few miles north of the Great Piece Meadows near the Pompton River in Wayne Township. Because of its proximity, the Packanack Lake core assumed to have kept a sediment mercury concentration record within the past century that is similar to that of the Great Piece Meadows, with concentrations ranging from 230-660 ng/g. Although Van Metre and Callender (1997) date the peak in concentration to 1940, more recent studies date it to ~1960 (Long et al., 2003; Van Metre and Callender, 1997)

<u>Mercury Background Concentrations</u> Background concentrations were established in order to distinguish between a rise in mercury concentrations due to coal burning and natural occurring concentrations within the surficial sediments. Organics within core TZS2 at a depth of 48 cm were dated to 5960 +/- 30 years before present (1950) so it can be assumed that mercury concentrations at the bottom of TZS1A, are background concentrations. Samples from the bottom 5 centimeters of the sampled

section of the core (42-38 cm) have a range in concentrations from 34.5-80.6 ng/g. Because of the low sedimentation rate implied by a radiocarbon date of nearly 6000 years at a depth of only 48 cm, concentrations from the bottom as well as the center of the core (~21cm) can also be considered background concentrations with a range of 34-147 ng/g. Organic matter from 50-54 cm of core RC2 were radiocarbon dated to 3260 +/- 30 years before present. It can also be assumed that samples from the bottom 10 centimeters of core RC1 (bottom 5 cm of RC2), which have a mercury concentration range of 34.6-79.6-ng/g, can also be considered background. OX1A was radiocarbon dated from a depth of about 49 cm to 650 +/- 30 years before present. Although the first 40 cm of the core was dated within the past 700 years, it is still likely that the bottom 10 centimeters of the sampled section of the core were deposited before major colonization of the area, and that background concentrations range from 90.8-168.7 ng/g. Total mercury background concentrations, taken from a table from Kroenke et al., 2002 list background measurements from different studies. Minnesota, Northern WI, has a background of 21-106 ng/g (Engstrom et al., 2007) background values fall within the range of the background values of the cores from the Great Piece meadows. Background mercury concentration levels for source rocks are determined to be 400 ng/g (Turekian and Wedepohl, 1961) to 180 ng/g (Marowsky and Wedepohl, 1971) for shales, and 30 ng/g (Turekian and Wedepohl, 1961) to 8-290 (Marowsky and Wedepohl, 1971) for sandstones. The majority of sediments within the cores are silts and clays, which may account for background values at the higher end of the ranges given by other studies.

Core TZS2

The oxbow meander cutoff from which core TZS2 was taken was determined by radiocarbon ages of 5960 +/- 30 and 7580 +/- 30 BP to be the oldest of the three oxbows, predating the other oxbows by at least 4,000 years. Its long record of silt deposits, poor LiDar definition, and its peat-like sediments further confirm the age of Oxbow TZS relative to the other oxbows. Although it follows the general fining upward sequence of the other cores, core TZS2 is unique in its stratigraphy in its large sand deposits and lack of visible stratified flood deposits. The approximately 87 cm of sand that makes up the lower section of Core TZS2 is interpreted to be fluvial channel deposits. The bottom ~42 cm of sand ranges from subangular medium-coarse sand to pebbles, which is coarser then the medium-fine sand collected off the banks of the modern river channel, and significantly coarser than any sand deposits found in the other two cores. The reason for this is possibly Core TZS2's proximity to the Rahway Till (Stanford, 2007b). The Rahway Till is described as a silty-sand to sandy-silt containing some to many subangular pebbles and cobbles (Stanford, 2007b). Both the current course of the river and previous paths shown by LiDar flow directly alongside the foot of Hook Mountain. Erosion of these sediments could have resulted in the deposition of coarser, more angular material in the TZS oxbow meander less than a mile downstream. Because of the Passaic's slow moving, gently sloping path through the Great Piece Meadows larger grained sediments would have been quickly deposited, and unable to make it further downstream to the locations of the other two oxbows. The shift in Core TZS2 to fine-medium sand around 168 cm may signify a change in river's course, and a decrease in glacial sediment as source material. The fining upward between 140-116 cm resulted from the meander gradually becoming closed off from the main river channel. The earliest radiocarbon date after the oxbow was cut off was 7580 +/- 30 years BP, obtained from peat-like organics at 74 cm depth.

Based on the amount of deposition between this date and the date at 48 cm, the meander is roughly estimated to have been completely cut off between 9,000 and 9,500 years BP. This may have occurred during a period of increased storm related floods that affected New England at around 9.1 ka (Noren et al., 2002). The top 103 cm of the core is silt, with little variation in color besides a lighter section around 27-63 cm. This is interpreted as either shallow lake or over bank floodplain deposits. Grain size results around 65 cm showed a slight increase in sand, which may be flood overbank deposits, but the thin layers of sand and organics that marked major floods in the other two oxbows are not seen in this core. One possible reason for this is the distance from the main channel. Two pieces of wood at 48 cm were dated to about 5960 years +/- 30 B.P. This indicates either that only about half a meter of sediment was deposited within 1,500 years, or core TZS2 underwent a large degree of compaction. The possible low amount of sedimentation may be due to avulsion (Kraus and Aslan, 1999). Low sedimentation suggests that the oxbow was completely cut off during avulsion and quickly shifted away from main channel, heavily restricting its sediment supply from the river.

Organic deposits in TZS2 also differ from the other cores. Silt from 105-65 cm is organic rich and similar to peat, varying from organic deposits of the other two cores, which either occur in layers or consist of scattered large wood, charcoal, or leaf fragments. (Fig. 23B). These heavily organic sediments suggest a saturated, but low energy environment that allows for regular local organic deposition, but does not receive many high energy flood deposits (Baker, 1987). Coarse channel sand, which the other oxbows lack, likely aided in point bar development, separating the oxbow from the rest of the channel, and reducing the oxbow's potential to be temporarily reactivated during major flooding. Another likely cause of a lack in visible organic deposit layers is the age of the core, which dates back past 8,000 years. Flood deposit layers may once have been easily distinguished, but have become obscured over time.

Unlike the other two oxbows, TZS does not contain a top clayey-silt layer. There is a rise in average mercury concentrations around 25 cm, but this is too deep to have occurred in the mid 1800s. Sediments around 4-5 cm show a drastic rise in mercury concentrations leading to a peak occurring around the 1950s, while sediments in the preceding 20 cm have concentrations fluctuating between 100-148 ng/g. This is a similar pattern to the Wawayanda mercury profile, implying that sediments in between 6 and 7 cm can be dated to about 1835, indicating about 6,000 years of deposition in only 43 cm. Mercury concentrations only peak in the top centimeter, suggesting either very low sedimentation rates since 1950-1960, or erosion. Erosion is suggested by the mercury concentrations, which peak at the top centimeter of the core, instead of peaking and receding before the surface of the core, due to recent emission regulations. However, the missing sediment record would be minimal and it is unlikely core TZS2 once contained a missing clayey-silt layer. During sampling, Oxbow TZS appeared to be the driest of the three oxbows. Lack of standing water and a dry upper layer of soil may have affected the deposition of clayey-silt deposits.

Oxbow2 and RC2 Cores

Based on the deepest radiocarbon dates of 3600 +/- 30 BP for Oxbow2 and 3560 +/- 30 BP for RC2, it appears that the two oxbows were contemporary river channel meanders. Wood at 178 cm for core RC2 and sticks, bark, charcoal, leaves and seeds at 197 cm for core Oxbow2, both within a silty-sand layer, were both dated to ~3600 +/- 30 years BP. Sediments around 200 cm depth for core Oxbow2 and 180 cm for core RC2 are considered to have been deposited during the transition from channel to oxbow cutoff. The upper section of the underlying glacial lake silt and clay of was only reached in core RC2, but surficial maps and test borings suggest that the clay would be found at about 2.5-6 meters below the surface in core Oxbow2 (Stanford, 2007b) The silty-sand and fine-very fine sand layers from 199-177 cm in core RC2 are interpreted to be channel deposits, though they are finer grained than the sand sample collected from the modern day river. However, a thin layer of coarse-very coarse sand was found in core RC2 within the transitional silty-sand layer between the fine sand and glacial silt and clay. These channel sediments are much finer grained and a significantly smaller channel sand unit than that of core TZS2.

Because the glacial lake sediments were not sampled within the core, it is difficult to tell how long the Oxbow2 meander remained an active part of the river. The bottom ~10 cm of core Oxbow2 contains medium to fine sand, which is the same grain size as sediments within the current river channel. While there is a layer of silt at 185-180 cm, the core is mainly comprised of silty-sand until about 103 cm, and doesn’t fully transition to fine grained lake or overbank deposits until ~65 cm. Core Oxbow2 is the core taken nearest to the river. The long record of sandy-silt, interspersed with thin layers of fine sand suggest that the meander was not immediately divided from the main channel, but rather gradually cut completely over a ~1200-1500 year span, frequently receiving channel sediments during periods of heavy flooding. The thin layers of coarser material surrounded by finer sediments can be interpreted as major flood events (Oliva et al., 2016).

The isolation of Oxbow2 from the main channel the river started around 3600 years B.P., dated from a possible flood deposit. About 14 major flood deposits are identifiable within the core. Deposits of lesser floods are assumed to also have been recorded, but are difficult to distinguish. If the radiocarbon date is correct, this removal from the main river channel occurred at the end of a Holocene dry period from 5-3.5 ka (Noren et al., 2002). Silt from 180-185 cm indicates it was temporarily cut off from the river before being sporadically activated again by flooding. Wood, charcoal, and seeds at 176 cm were dated to 3977 +/- 32 years B.P., which is older than the organic matter dated at 196.75 cm. Because the date of this organic matter is not in sequence, these organics are interpreted to have been washed in from an older source as flood deposits. Grain size results from 179-163 cm within the core fluctuated between sandy-silt to fine-medium sand. Images of this section of the core reveal poorly defined, infrequently spaced bands of 10YR 4/3 dull yellowish-brown and 10 YR 4/1 dark grey sediment, indicating that the changes in grain size are from influxes of sand rich sediment into the surrounding silt (Munsell, 1994). Though the sediment is mixed, four or five discernable sandy layers can be seen; which are likely high energy flood events that temporarily reactivated the oxbow and washed in sand from the adjacent active channel (Fig. 24C). Two thin layers of organics at 153 and 154 cm and one thicker layer at 147-150 cm are also considered to be flood deposits, especially since the material dated around 148 cm is older than the date at ~197 cm, indicating that older material was washed into the oxbow. Fine sand layers at 135, 116, and 109 cm are also considered flood deposits from higher energy floods. The influx of fine sand either indicates that floodwaters were able to carry the sand to the oxbow, or that the main river channel temporarily flowed into the oxbow. The oxbow was completely cut off from the main channel around 2350-2400 years B.P., based on a radiocarbon date from a centimeter thick layer wood, charcoal, and a small amount of seeds at 78.25-81 cm. This woody layer may correspond to a wood layer at ~49 cm in core OX1. OX1 also contains another wood layer around 57 cm, which Oxbow2 lacks, though it shows contains wood and charcoal deposits around 71 cm. These organics may be from the same event that deposited the wood layer in OX1, but flood deposits and sedimentation may have varied throughout the oxbow. It is impossible to totally correlate the two cores without more radiocarbon dates. Two grayish brown (10YR 5/2) nodules, which are lighter than the surrounding very dark greyish brown (10 YR 3/2) silt, are seen from 53-65 cm. Because of their shape and their similarity in color to the overlying layer, these nodules be a result of bioturbation, or pedogenesis, which can be seen as an indicator of a lower energy environment with little river influence (Fig. 24B). Flood deposits became less defined as the oxbow became completely removed from the river. Any flood deposits in the upper 65 cm of the core are silty overbank deposits, although there is some variation in color and charcoal content.

Core RC lacks the sandy high energy flood deposits of core Oxbow2 since it is located further from the main river channel, but it does include layers of organics that can be used as major flood indicators. Oxbow RC began to close off from the main channel between 3560 +/- 30 years B.P. and 3450 +/- 30 years B.P., after which sediment deposits consisted only of silt. This may have also coincided with the end of a Holocene dry period. Organic deposits occur in layers of seeds, sticks, leaves, and wood pieces, which are sometimes mixed in thick layers of unconsolidated silt are considered to be major flood deposits, eight which occur during a ~300 year span. A 3 cm thick deposit of wood at 108.5-105.5 cm dated to ~3420 +/- 30 years B.P. is thought to be from a driftwood deposits from a longstanding flood, which gave the flood transported organic material time to settle out. Although these organic flood deposits occur within the same time frame as the Oxbow2 core's, differences in sedimentation and flood deposit types make them difficult to correlate exactly.

Only the RC and Oxbow cores are topped with clayey-silt at about 25 and 7 cm, respectively. The change from silt to clayey-silt at around 25-30 cm is marked in core RC by a thin mud drape-type feature with less consolidated sediment, accompanied by mixing with pieces of lighter sediment, and possibly some charcoal (Fig. 25A). These features are not seen in the Oxbow core, but the core does exhibit an alternating light and dark banded layer from about 27-18 cm (Fig. 24A). Though difficult to determine because dates are estimated, mercury concentrations from an adjacent core, core RC1A, suggested that the deposits from the 1800's to present are preserved within the upper 20 cm of core. Core RC1A has lower peak and average mercury concentrations than most lakes in Northern NJ, although it does fall within the range of mercury measurements in other lakes. From previous studies, it is assumed that the core's peak concentrations can be dated to about 1950-1960 at 2 cm depth in RC1A. Because of the age of organics dated deeper in the core, sedimentation rates appear to be low; even less than that of Wawayanda Lake (0.1 cm/yr). The onset of widespread coal burning around 1835 is therefore estimated to have occurred around 12 cm, where concentrations begin to rise from the 35-42 ng/g values of the previous 10 cm to values of 50-75 ng/g before peaking. This marker occurs at about 7 cm within core RC2A because of the 5 cm difference between RC1A and RC2A. This suggests that Oxbow RC experienced a drastic decrease in sediment deposition in the upper section of the core, going from about 70-86 cm of deposition in about 100 years according to the radiocarbon dates, to over 3,000 years of deposition within the upper 50 cm of the core.

Radiocarbon dating in the sample core taken on the downstream side of Oxbow2 (OX1) give seeds at about 49 cm a date of 650 +/- 30 years B.P., while a mercury concentration peak at 10 cm dates to about 1950. Core OX1A first peaks at 10 cm and peaks again at 7 cm, likely spanning from 1940-1960. Coal powered industrialization is estimated to occur around 17-18 cm within core OX1A. If both these dates are accurate, sediment accumulation within the upper 50 cm of core ranges from 10-15 years per centimeter, assuming the top of the core can be dated to the present year. This is just an estimate, since the true dates that coincide with the mercury peaks are unknown. However, despite surface variations across the floodplain, it appears that sediment deposited since the early 1800s is contained within the upper 20 cm of floodplain stratigraphy. Any deposits from recorded floods will be present 20 cm or less below the surface.

The long record of silt deposits in core RC2 possibly means that Oxbow RC2 was more developed as an oxbow lake than the other two oxbows. While it did not receive coarse grained flood deposits from the active channel, it was able to achieve a fast, steady sedimentation rate as a shallow lake environment that was able to preserve a record of flood deposits. A loss of identifiable flood deposits is seen in the upper ~45 of both the Oxbow2 and RC2 cores. Though the upper sections of the cores lack the well-defined sand and organic layers at greater depths, they aren't devoid of possible flood deposits. Core RC2A shows a mud drape type feature at ~30 cm depth (35 cm in RC1A) and a less consolidated darker layer at 25 cm. Core Oxbow2 has charcoal within the upper 50 cm of core, and banded or mottled sediment around 15-21 cm 47-53 cm, which could be indicative of flood deposits. Flood deposits in the upper sections of the core may also depend on location, since core OX1 shows a thin organic layer around 30 cm, and organics and wood around 43-53 cm. The difference in flood deposits and rate of sedimentation between the upper sections of cores Oxbow2 and RC2 and the lower sections is certainly influenced by of the river moving farther away from the oxbow, and the draining of oxbow lakes. A lake environment would have preserved a better sedimentation record than the current wetland environment. Though the change in flood deposits appears to have started before the area was widely settled, there may be an anthropogenic factor as well. Contrary to this study, research on prehistoric settlement in New Jersey around 1000-1300 A.D. show an increase in sedimentation because of deforestation for farming (Stinchcomb et al., 2011). However, drainage of the oxbows may have increased after European settlement when the wetlands were drained for farming (Brydon, 1974). Another anthropogenic alteration to the drainage basin, the construction of reservoirs and dams, may also affect more recent sediment supply by trapping river sediments upstream along the Passaic and its tributaries. Large woody debris is also decreased in streams in urban areas (Finkenbine et al., 2000). Changes in flood deposits and sedimentation may also signify a decrease in the river's ability to move sediment and debris. This is possibly because of a change in the river's gradient, or change in regional climate.

While the river's low gradient makes avulsion an anomaly, several aspects of the basin's history may explain the formation of the oxbows and their changes in flood deposits since the Holocene. Isostatic rebound has caused a gradient of 0.4 m/km parallel to the Northeast flowing section of the Passaic River (Stanford et al., 2016; Stone et al., 1989). The area around the Great Piece Meadows has experienced a rise of about 12-16m since the start of rebound in this area at about 14 ka B.P. (Engelhart et al., 2009; Koteff et al., 1993; Stanford et al., 2016). Calculated rebound half-lives of New England are about 1,000-1,250 years (Dyke and Peltier, 2000). Based on these values, the Central Basin experienced about 7.86 mm- 3.14 cm of rebound from 4- ~2.75 or 3 ka B.P. and only about 7 mm – 2.4 cm of rebound from about 3 ka B.P. to the turn of the century. This change in rebound may not have been enough to alter the gradient of the river and change major flood sedimentation, but it may have still had an effect. Changes in gradient around the time the Oxbow2 and RC oxbows began to be cut off from the main channel (~3.5-4 ka B.P.) only amount to ~5.86 mm – 6.3 cm, and would be unlikely to have caused river avulsion. However, 0.375- 1.001 m of rebound occurred from ~10.25-9 ka B.P., and 0.188- 0.503 m from ~9-7.75 ka B.P. Though still a small change in gradient, this rebound may have affected the cutoff of oxbow TZS, which occurred within these time periods, and the change from sand to silt deposits within the oxbow. A more likely possible cause for the change in flood deposits is climate. Lacustrine records of Lake

Blauvelt in Franklin Lakes, New Jersey show wetter conditions and increases in mean grain size and total organics from about 10,000-8,500 years BP and about 4,000-250 years BP; coinciding with the estimated removals of all three oxbows from the main channel (Getch et al., 2016). The cutoff of the Oxbow2 and RC2 meanders also occurred after a long dry period at ~5-3.5 ka, while the proposed cutoff of oxbow TZS follows North Atlantic Ocean recorded cold events during the Holocene as well as an increased period of storm related floods in New England (Getch et al., 2016; Li et al., 2006; Noren et al., 2002)(Fig. 27). The change from a dry to a wet climate period may have been the reason for the river's rare avulsions that caused the oxbows. A period of increased precipitation, especially after a long dry period, would have cause increased streamflows within the Passaic and its tributaries. A long period of dry conditions would have reduced soil moisture and dense vegetation within the basin, leading to a decrease in soil stability. Flash floods may have caused the avulsion episodes that formed the oxbows. Other changes in regional climate after 2400 years B.P., such as the cold period recorded by White Lake around 1.3 ka, may have affected flood sedimentation for the Passaic (Li et al., 2006) Though further study is needed, the effect of climate changes on the Passaic and its tributaries may also be the reason for the change in major flood deposits.

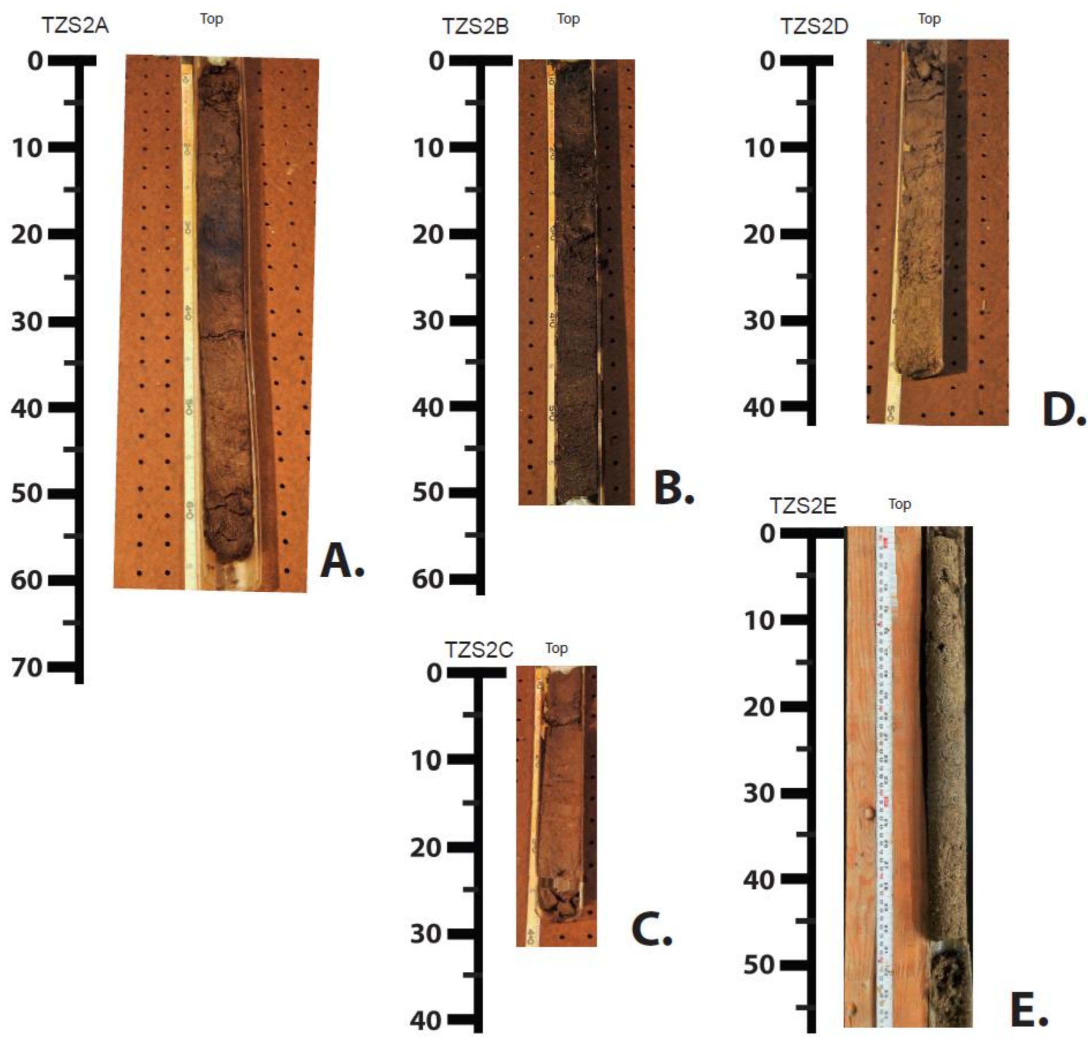


Fig. 23 A-E: The five sections of core TZS2.
Scale is in centimeters.

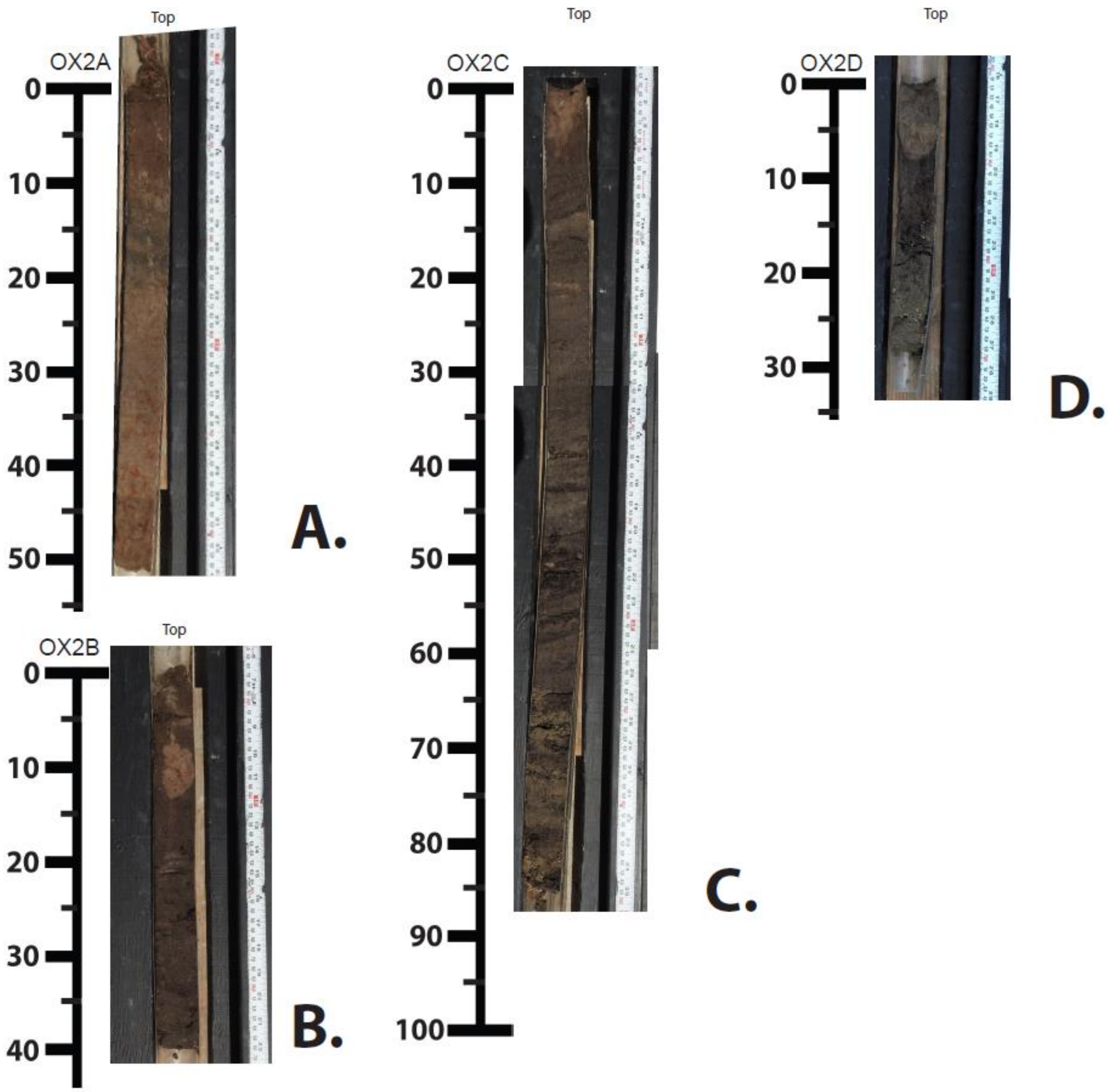


Fig. 24 A-D: The three sections of core Oxbow2. Scale is in centimeters

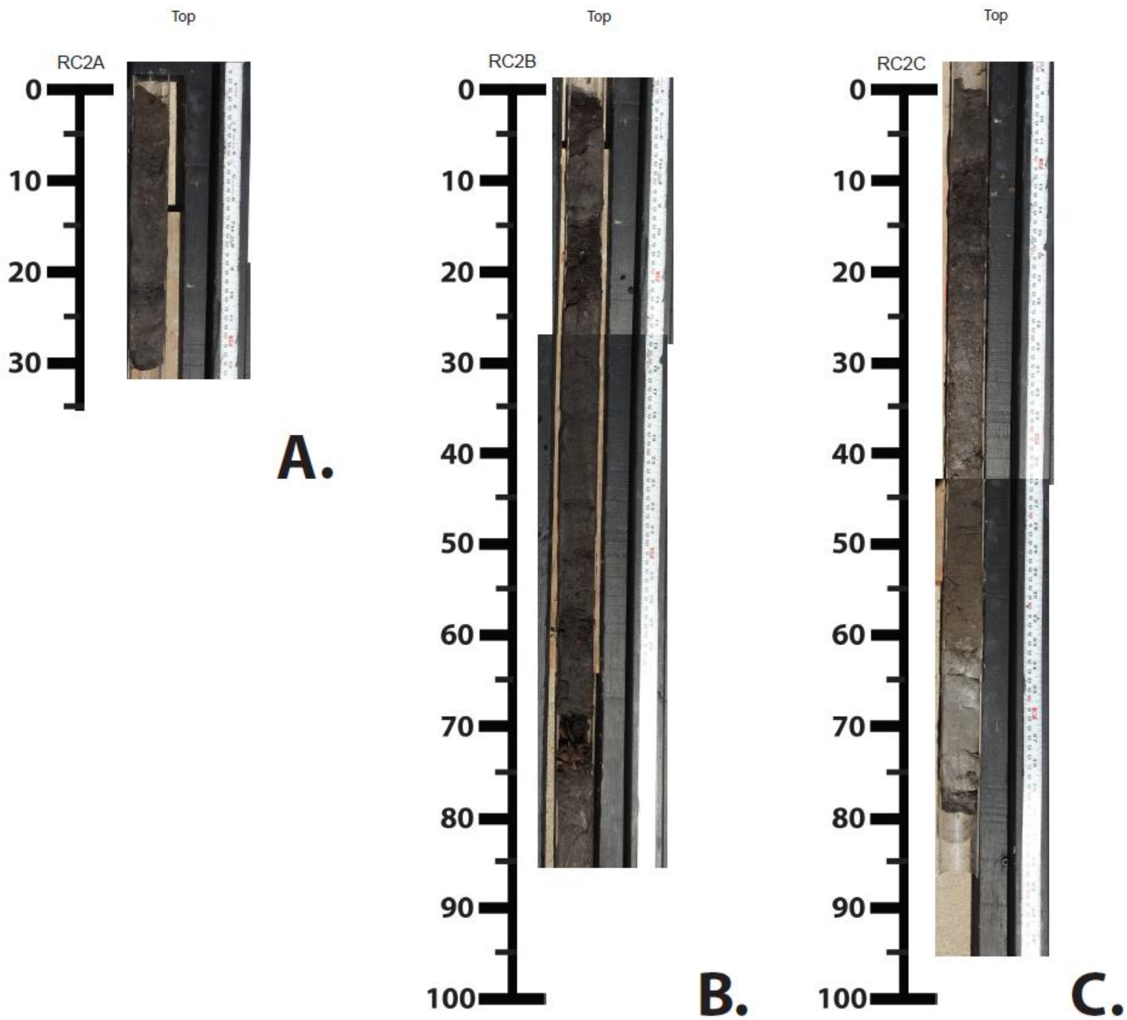


Fig. 25 A-C: The three sections of core RC2. Scale is in centimeters

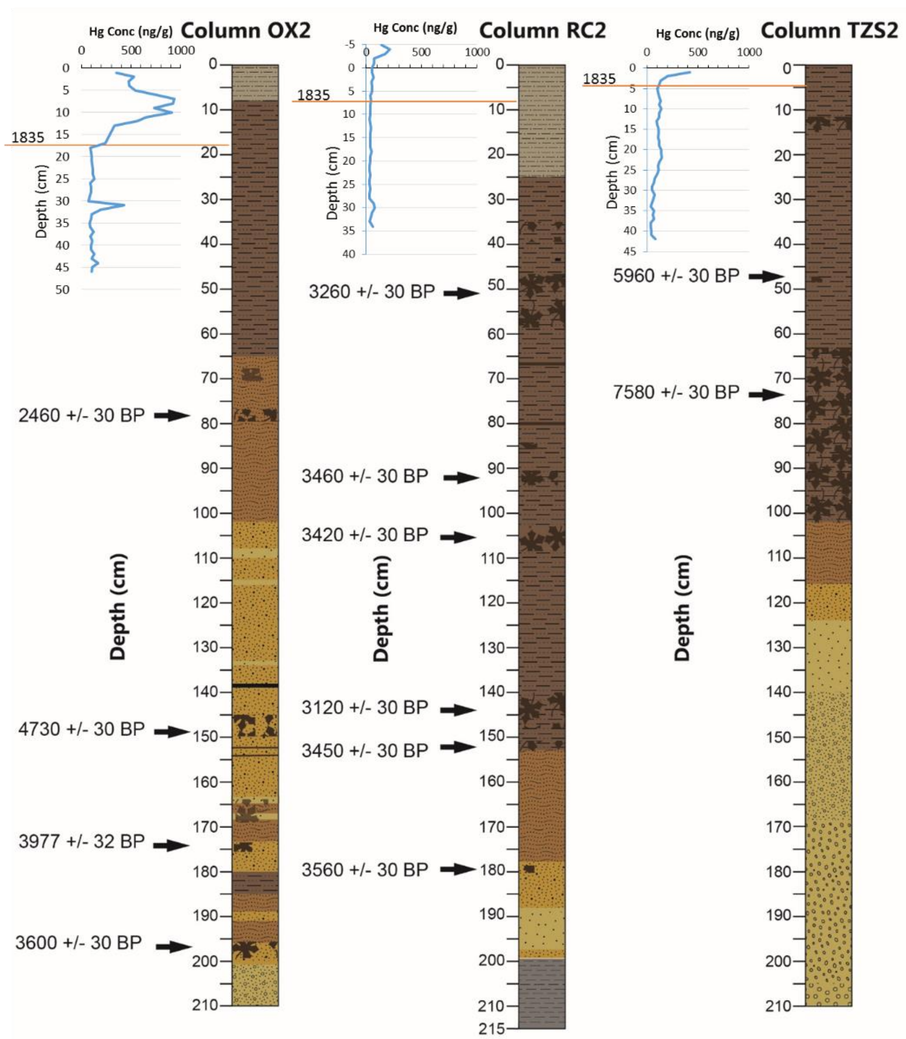


Fig. 26: Mercury concentrations relative to core depth for cores Oxbow2, RC2, and TZS2

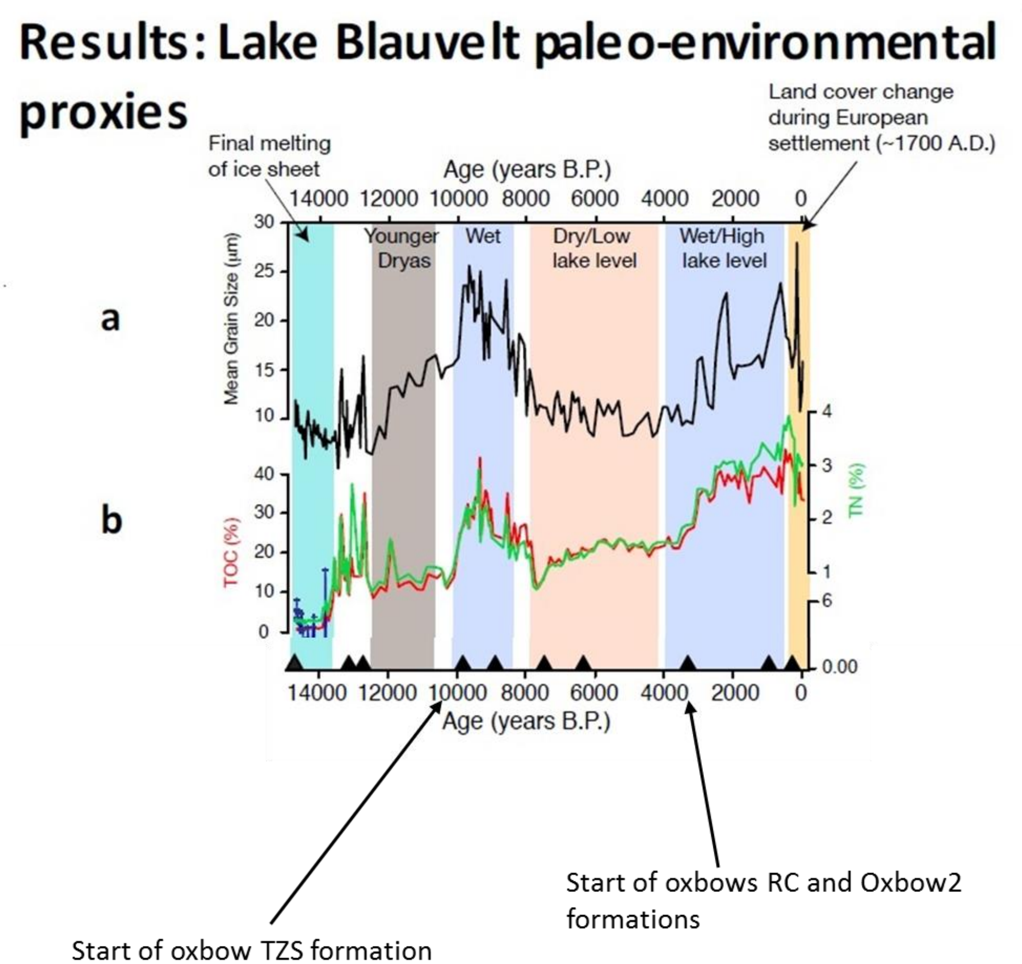


Fig. 27: Study of lake records from Lake Blauvelt in Northern NJ showing mean grain size and total organic carbon (TOC) fluctuations corresponding to age (years B.P.). Modified from (Getch et al., 2016). The approximate occurrence of the oxbow avulsions are labeled in the timeline

HYDROLOGY AND FLOODS

Hydrology and Flooding History

Climate A rise in precipitation since the end of the 1800's is a probable cause for increases in flooding frequency and intensity within the basin. Precipitation has had a gradual increasing trend since 1895, and an increasing number of years with precipitation higher than the historical average occurring since the 1970s (Office of the NJ State Climatologist, 1994-2017a; Southern Climate Impacts Planning Program, 2017). Five year moving averages of average annual precipitation also show an increase since the turn of the century, although a decrease does occur between 1895 and 1970. Climate data for Division 1 in New Jersey from the Southern Climate Impacts Planning Program shows two major peaks spanning 10 years or greater and three minor peaks spanning 2-4 years of annual precipitation greater than the historical average of 47 inches since 1970. The highest average annual precipitation on record occurred in 2011 with 72.49 inches of precipitation, 25.46 inches higher than the historical average. The last major peak occurred from 1895-1906. A series of precipitation peaks lower than the historical average occurred from 1905-1936, and from 1955-1970 (Southern Climate Impacts Planning Program, 2017). Average annual temperatures have risen from 1895 to present, with a sharp rise since 1980 in which all annual temperatures greater than the historical average of 49.6 °F (Southern Climate Impacts Planning Program, 2017).

Peak Discharge at Little Falls Gage Since 1870, there have been at least 37 major floods in which the peak discharge was at 10,000 cfs or greater recorded by the Passaic River gage at Little Falls (Fig. 28)(Hollister and Leighton, 1903; Leighton, 1904; Salisbury and Knapp, 1897; Smock, 1891; USGS, 2017; Vermeule, 1894) . Records of major floods at the Great Falls gage downstream suggest there may have been about two or three more major floods between 1870 and 1900, but records for these floods could not be found for Little Falls. Peak discharge for major floods has varied throughout the record of the Passaic River at Little Falls gage. At first glance, major flooding issues appear to be getting better since the late 1800s. Peak major flood discharge values after the flood of 1936 are significantly lower than those in the late 1800s and early 1900s (Hollister and Leighton, 1903; Leighton, 1904; Salisbury and Knapp, 1897; Smock, 1891; USGS, 2017). While the flood of 1903 is a rare flood event that is the highest on record, the flood of 1902 and recent preceding floods of the late 1800s also have peak values greater than those of most floods after 1940 (Fig. 28) (Hollister and Leighton, 1903; Leighton, 1904; USGS, 2017). Historical records also show frequent major floods occurring in the late 1800s (Hofman, 1955). The decrease in peak flood discharges can be attributed to the construction of major reservoirs in the late 1800s and 1920s, and increasing diversions from the Passaic and its tributaries for water supply (NJGS, 2013; USGS, 2005). Although the average peak discharge during major flooding remains lower than those of the 1800s and early 1900s, major floods appear to be becoming more frequent for the central Passaic. Flood frequency increased after the drought in the 1960’s, major flood frequency per decade similar or higher than during 1870-1910. After the 1960’s, three or more floods would occur within the same decade, with multiple floods sometimes occurring within the same year. Some floods, such as the flood of April 1984 and August 2011 had peak discharges comparable to those at the turn of the century (Fig. 28). Increased frequency may in part due to rising annual precipitation and temperature averages. Frequent flooding in the 1970’s coincides with years of high annual precipitation. The flood caused by hurricane Irene in 2011 occurred during the year of the highest average annual precipitation on record. Though average peak flood discharge has decreased since the turn of the century, the frequency and growing intensity of major floods are increasing flooding issues for the Passaic River.

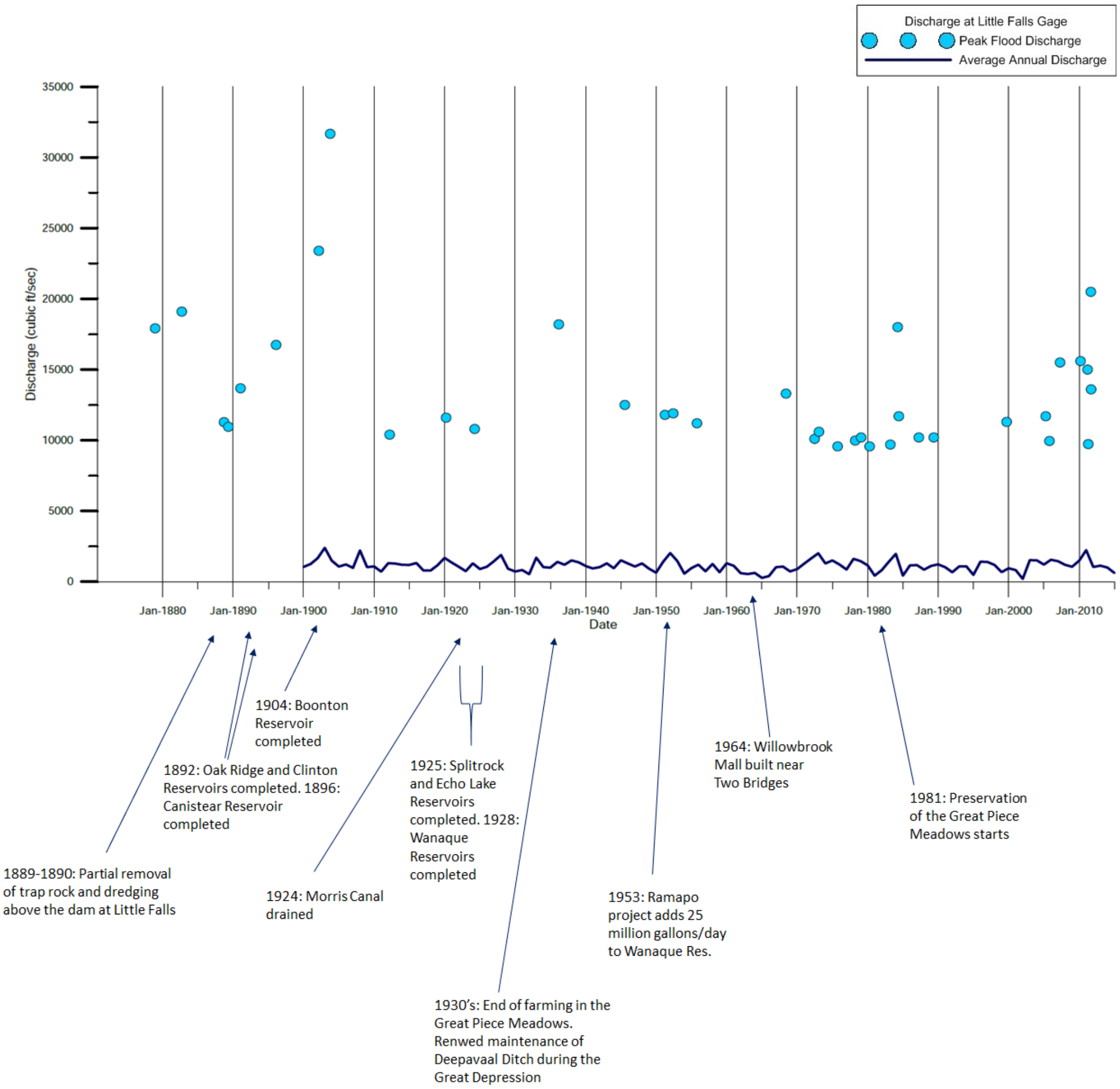


Fig. 28: Peak flows of major floods at the Passaic River at Little Falls gage. Data is divided into 10 year intervals. Data from (Hollister and Leighton, 1903; Leighton, 1904; Salisbury and Knapp, 1897; Smock, 1891; USGS, 2017; Vermeule, 1894).

Runoff Ratio Runoff ratio values were calculated for the drainage area of each stream gage to remove the effect of increasing mean annual precipitation (Office of the NJ State Climatologist, 1994-2017a; USGS, 2017). It allows for only observation of only the ratio of precipitation that contributes to mean annual streamflow through runoff instead of percolating down through the surface (Fig. 15). This allows examination of the effect of increasing impervious surfaces due to development within the basin on river discharge. All stream gages except for the Whippany River near Pine Brook (Fig. 29F), Wanaque River at Wanaque (Fig. 29B), and Passaic River at Little Falls (Fig. 29A) gages showed an increasing trend in runoff ratio values over the span of their records, which can be linked to water supply operations within the drainage basin. Out of the three gages with a decreasing trend, Wanaque River at Wanaque showed the sharpest decrease, with values ranging from and 0.79-0.35 until the 1930s to 0.32-0.04 since 1985 (Fig. 29B). This sharp decrease is a result of the construction of the Wanaque Reservoir in ~1928, the largest of the reservoirs and located just upstream of the gage, and an increase in diversion in 1932 (Fig. 29)(NJGS, 2013). Both factors affected the steep drop in the annual runoff ratio around 1931. The decrease in the annual runoff ratio for the Whippany River near Pine Brook gage is a result of a high peak ratio greater than one at the beginning of its short record in 1997. The other Whippany River gage upstream at Morristown (Fig. 29L) has a longer record, and shows an increasing trend in the runoff ratio with a high peak of 0.89 in 1997. It is most likely that the high runoff ratios are caused by effluent discharges from nearby sewer treatment plants.

The Rockaway River below Boonton Reservoir gage (Fig. 29J) also has a runoff ratio value greater than one for 1908. The difference between the values for local and regional annual precipitation may be one of the causes of the abnormal runoff ratio value. The average annual precipitation for the town of Boonton is 46.26 for 1908; over two inches greater than precipitation for Northern NJ (Office of the NJ State Climatologist Historical Monthly Station et al., 1994-2017). The Passaic River at Little Falls gage also has a decreasing runoff ratio trend, likely from the construction of the reservoirs and water diversions from the Passaic and its tributaries. Diversions by the Passaic Valley Water Commission (PVWC) have continued since the turn of the century (Fig. 30). Although diversions from the Passaic by the PVWC have a general increasing trend to 60 MG/day or greater, there have been fluctuations over the past century (NJGS, 2013). Low diversions less than 10 mg/day occurred between 1932 and 1942. The years 1999 and 2007 included months during which there was zero withdrawal, but diversions from the Pompton River by the PVWC were started around this time (NJGS, 2013).

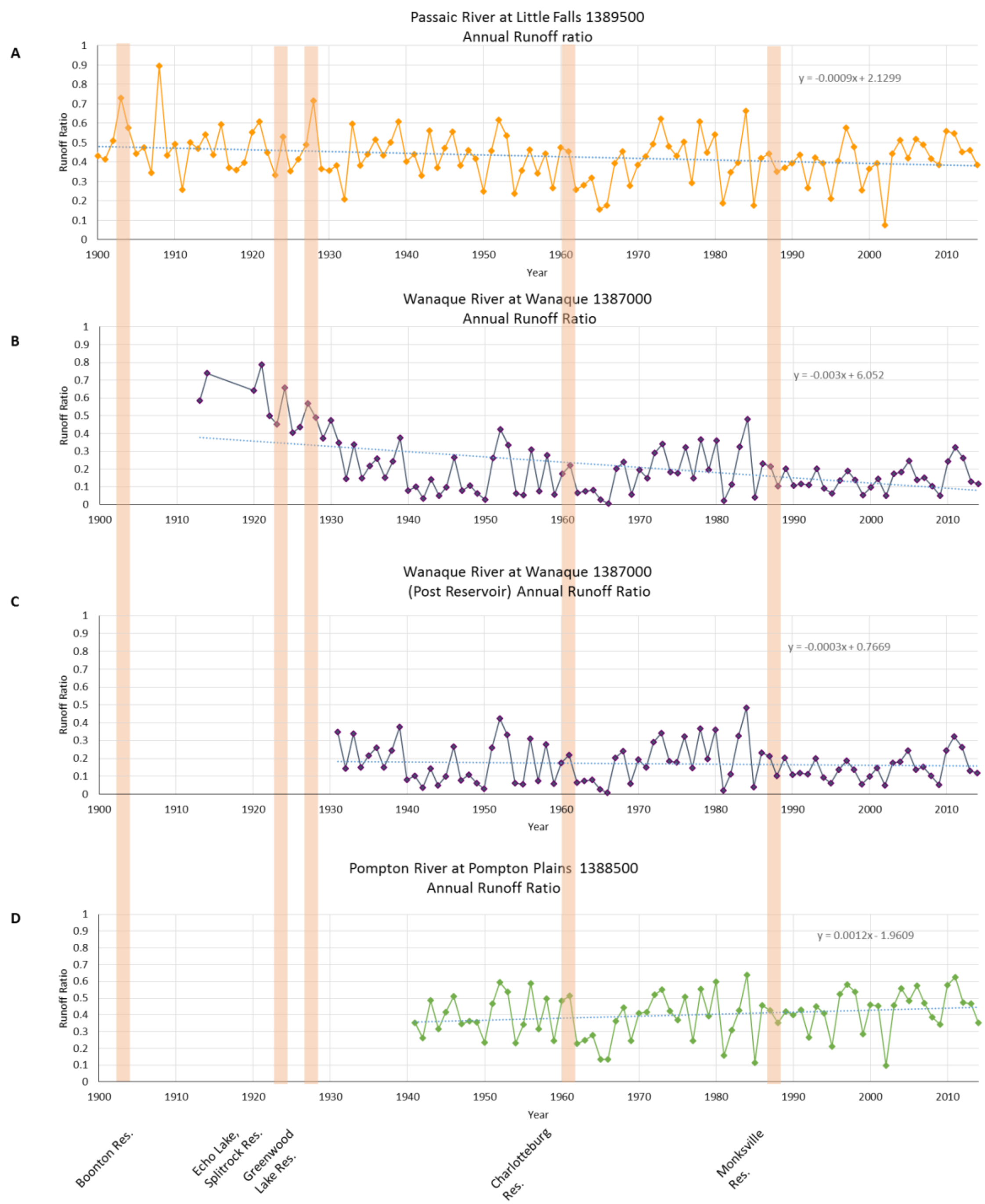
Despite having less impervious surfaces, gages in more rural areas did not exhibit the lowest runoff ratios. The Ramapo River gage had the steepest rise in runoff ratio. However, this is likely due to its short record, which started in the 1980s. The Ramapo River at Ramapo (Fig. 29K) gage is situated in the most rural area, compared to the heavily populated areas around the other gages, but does not have the lowest annual runoff ratios. The gages with the lowest runoff ratios, Wanaque River at Wanaque, Pequannock River below Macopin Intake Dam (Fig. 29H), and Rockaway River below Boonton Reservoir, are all directly downstream of a facility that diverts or regulates streamflow, such as a reservoir or a dam. The gage on the Rockaway River above Boonton Reservoir has significantly higher runoff ratio values than the gage below the reservoir.

Local rainfall, the amount of developed land within the river drainage area, and reservoirs within the drainage area all affect discharge during flooding. While most of the years during which a major flood occurred show a peak or an increase in runoff ratio for each gage, high average annual runoff ratios do not always coincide with years with major floods. Two or more gages showed annual runoff ratio peaks in 1908 (Rockaway River below Boonton Reservoir and Passaic River at Little Falls) and 1997 (the Whippany River at Morristown and Pine Brook), which were not years of major floods. While the exact cause of the abnormally high value for the Rockaway River below Boonton Reservoir in 1908 unclear, an event such as the release of a large amount of water from the reservoir to the Rockaway River downstream or effluent from nearby water treatment facilities could increase the river's discharge despite annual precipitation not being abnormally high for that year. The Passaic River at Chatham gage had a runoff ratio peak for 1978, during which there was a major flood. Two gages had peaks in 1984 and 2011, which were both years that included two or more major floods. The year 2011 also had the highest average annual precipitation since 1885, and four major flood events (two floods during Hurricane Irene). Multiple major floods within the same year are reflected by high annual runoff ratios, though not all high annual runoff ratios can be related to years with major floods.

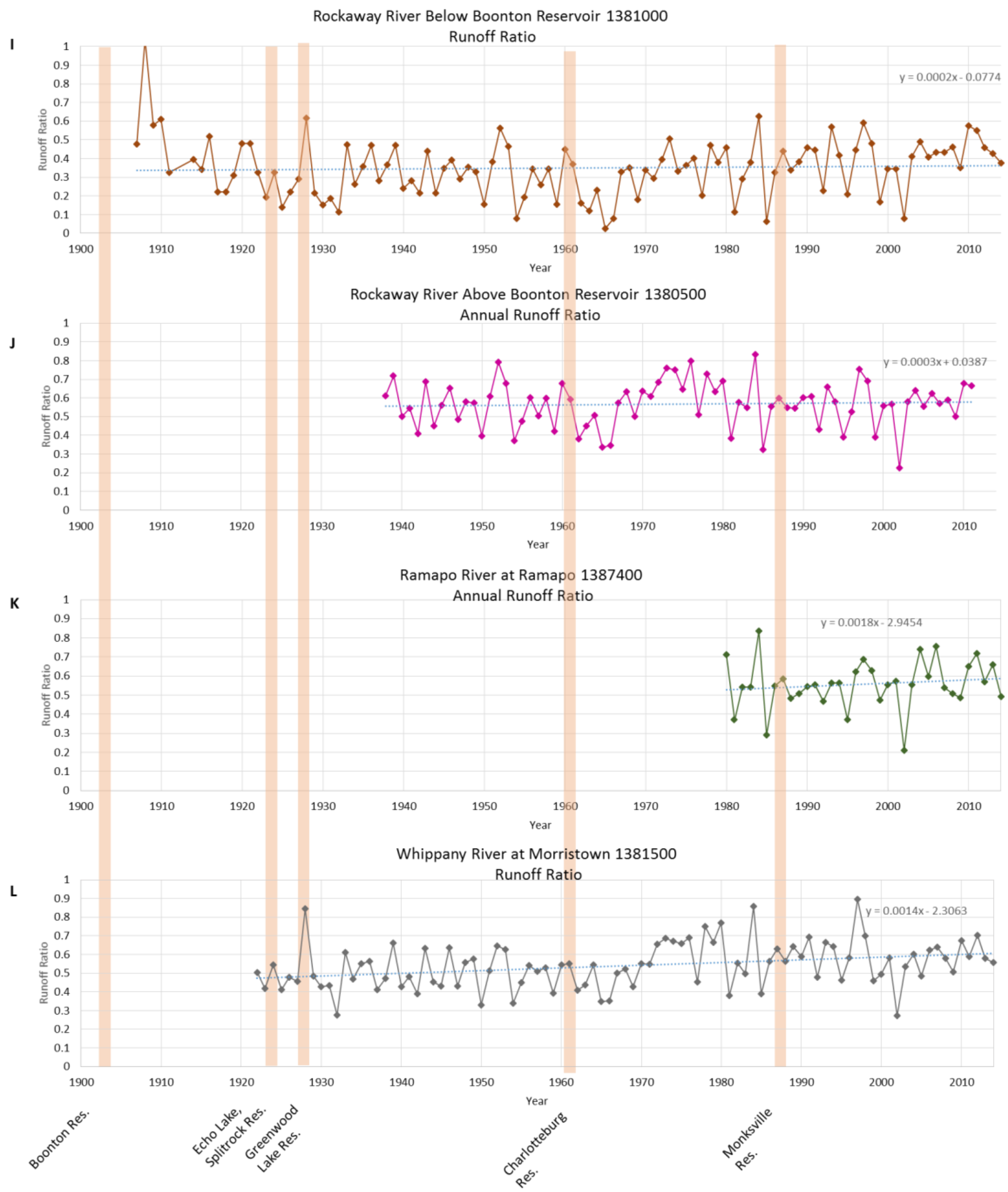
Most of the lowest annual runoff ratios for all gages occur during periods of drought for Northern New Jersey, both because of lack of rainfall and an increase in diversion to the reservoirs. The lowest runoff ratio value of the gages occurred either in 1965-1966 or 2002, with the exception of the Passaic River at Chatham gage, which occurs in 1911 (Bauersfeld and Schopp, 1991; Hoffman and Domber, 2004). The 1965-

1966 values occur during the worst drought on record, which lasted six years in the 1960s and included the lowest annual precipitation on record (Office of the NJ State Climatologist, 1994-2017a). While each gage shows a significant deficit in the runoff ratio during the mid 1960's, more gages have their lowest runoff ratio value during the drought of 2002, especially those gages located directly downstream of a reservoir or dam. While storms and floods impact the annual runoff ratio of the Passaic and its tributaries, the magnitude of their impacts varies among the rivers and their gages. Overall, droughts have the largest impact on annual runoff within the entirety of the Passaic River Basin, likely due to the increase in water storage and diversions for water supply (NJGS, 2013).

Fig. 29 A-L: Annual runoff ratios for the stream gages of the Passaic River Basin. Pink bars indicate years during which reservoirs were completed. Data from (Office of the NJ State Climatologist, 1994-2017a; USGS, 2017)




E
Passaic River near Millington 1379000
Annual Runoff Ratio
y = 0.0005x - 0.4925
Runoff Ratio
Year
F
Whippany River near Pine Brook 1381800
Annual Runoff Ratio
y = -0.0137x + 28.148
Runoff Ratio
Year
G
Passaic River near Chatham 1379500
Annual Runoff Ratio
y = 0.0006x - 0.7202
Runoff Ratio
Year
H
Pequannock River at Macopin Intake Dam 1382500
Annual Runoff Ratio
y = 0.0011x - 1.8467
Runoff Ratio
Year
Boonton Res.
Echo Lake, Splitrock Res.
Greenwood Lake Res.
Charlotteburg Res.
Monksville Res.



I
Rockaway River Below Boonton Reservoir 1381000
Runoff Ratio
y = 0.0002x - 0.0774
Runoff Ratio
Year
J
Rockaway River Above Boonton Reservoir 1380500
Annual Runoff Ratio
y = 0.0003x + 0.0387
Runoff Ratio
Year
K
Ramapo River at Ramapo 1387400
Annual Runoff Ratio
y = 0.0018x - 2.9454
Runoff Ratio
Year
L
Whippany River at Morristown 1381500
Runoff Ratio
y = 0.0014x - 2.3063
Runoff Ratio
Year
Boonton Res.
Echo Lake, Splitrock Res.
Greenwood Lake Res.
Charlotteburg Res.
Monksville Res.

<u>Flood Discharge Patterns</u> Analysis of five day peak flood discharges during major floods over the past century has yielded information about the behavior of the Passaic and its tributaries, as well as factors affecting flood discharge (Fig. 19). The discharge data confirms previous observations that the Upper Passaic and the tributaries of the Passaic and Pompton show a peak in flood discharge one or two days before the Central Passaic peaks in discharge at Little Falls (Fig. 1) (Hollister and Leighton, 1903; US Army Corps of Engineers, 1987). This is expected during flooding events, since the Little Falls gage is located a few miles downriver where the Pompton River joins the Passaic, so it receives the combined flows of the Passaic's tributaries as well as the Pompton's. The USGS discharge data is an average of the daily discharge, and so the difference in peak flood discharge between the Little Falls gage and the tributary gages may be several hours rather than a full day. However, the data still shows an overall trend. The Highland tributaries are at a higher elevation than the Passaic, so they are the first to receive precipitation that falls on the mountains, and waters from snowmelt in the spring. They also have a steeper gradient and narrower channels because of their underlying igneous and metamorphic bedrock geology compared to the low gradient Passaic, which is underlain by more easily eroded sedimentary bedrock (Drake et al., 1996; Hollister and Leighton, 1903). The Highland tributaries' geology and morphology cause them to reach their peak discharges quickly during storm events. However, there are some exceptions. During the floods of 1903 and 1945, both the Pompton River at Pompton Plains and Passaic River at Little Falls gage peaked on the same day, most likely because of the magnitude of flooding. During the second flooding event resulting from hurricane Irene in 2011, both the Rockaway River gages above and below Boonton Reservoir peaked on

the same day as the Passaic River at Little Falls gage, although they had peaked a day before the Little Falls gage during the first flood event a week earlier (USGS, 2017). In both 1936 and 1955, the Wanaque River at Wanaque gage peaked in discharge at the same time as the Little Falls gage, and so did the Rockaway River below Boonton Reservoir gage in 2005. Both of these gages are similar in that they are situated directly below reservoirs, which can act as buffers to floodwaters.

Flood discharge patterns also reflect the restrictions of flow at the gap in the mountains at Little Falls, and the subsequent flooding that occurs upstream. The Passaic gage at Little Falls usually records the highest peak discharges during major floods, surpassing the other gages by at least 7,000 cfs (USGS, 2017). However, the Pompton River at Pompton River Plains gage occasionally records higher peak flood discharges than at Little Falls, such as during the floods of 1886, 1955, April 1984, 1987, 1999, 2005, 2007, April 2011, August 2011, and March 2010. Both the Pompton River at Pompton Plains and Passaic River at Little Falls gages have similar gradients along the sections of the rivers that include their gage locations, at about 10 ft/mi (Spitz, 2007). Yet the Pompton River exhibits “flashy” discharge similar to that of the tributaries, with discharges sharply declining after flood peaks. Discharge at the Little Falls gage has a more gradual decrease after peak flood discharge, with discharge measurements still above half of peak discharge, two days later. This gradual reduction in discharge is likely because floodwaters back up around Little Falls and pool in the Great Piece Meadows instead of rapidly draining. Both of the Passaic gages upstream of Little Falls peak with the tributaries, but have fewer fluctuations in discharge. The difference in flood peak patterns between the tributaries the Central Passaic River also give evidence to natural causes of flooding. Water from the steep, quick spilling tributaries flowing into the flat, poorly draining Central Passaic Basin sets the area up for flooding (Hollister and Leighton, 1903).

The Ramapo River at Ramapo gage, which is located within a more rural area compared to the other gages, has the third highest peak discharges during flooding events, and occasionally peaks a day before the other tributaries; likely due to the gage location as the farthest upstream in the Passaic River Basin. Its discharge is that of a quick flowing mountain stream, and decreases more drastically than the other tributary gages besides the Pompton after peaking. The highest discharge recorded for this gage is during the flood of 1896 at 8731 cfs, even though the river was reported as being restrained by winter ice (Salisbury and Knapp, 1897). However, this peak discharge was almost reached during Hurricane Irene. While the high, flashy discharge exhibited by the Ramapo River at this gage is due to its elevation, its location may also be a factor. Located in a less developed area further upstream in the drainage basin than the other stream gages, it is not subjected to the water diversions, reservoirs, dams, and other constrictions that affect the flow of streams in more urban areas downstream.

For most of the floods, heavy precipitation over northern New Jersey (Division 1) during or prior to the month of the flood (especially during winter months) is the obvious cause (Fig. 18A). However, causes for major floods are not always clear. The flood of November 1977 had only about 7 inches of precipitation that month over the Division 1 area, and even less the previous month (Office of the NJ State Climatologist, 1994-2017a). However, the town of Little Falls received 11 inches of precipitation, causing major flooding in the Lower and Central Passaic River basin (Office of the NJ State

Climatologist, 1994-2017b). Yet for some floods both regional and local precipitation are not especially high, and flooding seems unlikely. Individual examination of flooding events is needed to fully understand the causes of major floods, but precipitation does not seem to be the only factor.

The flood of 1902 was a direct result of heavy precipitation, and is one of the worst floods in the past century. While 6 inches of precipitation is not usually high for the region, it it is high for the month of February (Office of the NJ State Climatologist, 1994-2017a). The 1902 flood occurred from Feb 25-March 9 1902, reaching a peak on March 2 (Hollister and Leighton, 1903). Heavy precipitation that exceeded normal standards had occurred during several consecutive storms that February. The melting of previous snow cover also added to the influx of water. The water began to rise in Chatham in the Upper Passaic River on Feb 27, reaching a peak on March 1. The highland tributaries gages all peaked about half a day before the gage at Little Falls. While the combined flow at the gages peaked more sharply than the flow at Little Falls, the decline in flow was slightly more gradual(USGS, 2017). The gages and bridges at Two Bridges were fully submerged, and all farmland in the area was completely inundated. The Lower Passaic was the most populated area along the Passaic in the early 1900s, and was heavily urbanized. Though heavy precipitation caused the flood, its large areas of paved roads providing “quick spilling surfaces”, contributing to the high amount of damage in the area (Hollister and Leighton, 1903).

The flood of 1903 was the most severe on record, and was caused by heavy rainfall of about 11.74 inches in the days prior to the flood (Leighton, 1904). The reservoirs were at high storage capacity during this time as well. On October 8, the

Passaic River overflowed its banks and remained flooded until October 19. Maximum levels were recorded at Dundee Dam at 9.5 feet over dam crest, Little Falls at 1.29 feet over the dam, and over 14.3 feet at the survey gage at the feeder of the Morris Canal in Pompton Plains. Flooding in the Central Basin was not as destructive as that in the Lower Basin because a lake formed within the Great Piece Meadows, buffering flows from higher elevations (Leighton, 1904). However, reservoirs and lakes within the drainage basin, which usually provide some storm water storage, were at high capacity during the heavy rainfall and therefore did not deter extreme flooding. The flood crest reached the Upper Passaic a day later than the Pompton and the Rockaway Rivers, and 12 hours later than the Lower Passaic (Leighton, 1904). However, both the Passaic River at Little Falls and the Pompton River at Pompton Plains gages peaked on the same day at 31,675 cfs and 28,300 cfs, respectively. These are the highest values of record.

The flood of May 1968 was caused by heavy precipitation for the month of May for the Division 1 region. Little Falls experienced even greater precipitation at about 9 inches for the month of May. This flood caused widespread damage around the Passaic and its tributaries, most severely affecting the Pompton, Wanaque, Pequannock, and Ramapo Rivers (US Army Corps of Engineers, 1987). Discharge along the Upper Passaic and the tributaries peaked a day prior to the Passaic River at Little Falls, and quickly dropped. Compared to another major flood in May, the flood of 1989, the Passaic River at Little Falls had a higher peak of ~13,000 cfs as opposed to 10,000 cfs, despite lower precipitation (Office of the NJ State Climatologist, 1994-2017a; USGS, 2017). The peaks of the tributaries of the two floods were similar showing that other factors besides precipitation affect flooding intensity differently among the tributaries.

While Northern New Jersey did experience unusually high precipitation during April 1984, previous snowfall from a coastal storm in March, melted by rain during an April storm produced the worst flooding in almost 50 years (Philips and Schopp, 1986). Above average precipitation from October 1983 to March 1984 left reservoirs and lakes at high capacity and the ground saturated before the coastal storm of March 30 and the storm of April 4-6. During the April storm, 2-8 inches of rain were deposited over Northeastern New Jersey. Flood crests of the Wanaque and Ramapo Rivers were greater than those predicted for a 100 year cycle (Philips and Schopp, 1986). The central section of the river basin, which had the extensive development on its floodplains, experienced the most damage. The upper section had mainly narrow channels, which allowed for only limited flooding. As shown in the hydrographs, the discharge at the Little Falls gage remained above 10,000 cfs at least two days after the peak flood discharge, while discharge at the Upper Passaic and tributary gages dropped near to normal levels three days after peaking (USGS, 2017). Both the Wanaque and Boonton Reservoirs were at high capacity the month before and during the flood, and so were not able to buffer flooding downstream (NJGS, 2013). The swamps and lowlands of the central basin provided storage that alleviated some of the flooding problems of the lower basin (Philips and Schopp, 1986)

While not much is written about the nature of the flood January 1979 it appears to be an example of a flood caused by heavy winter precipitation. High discharge values are seen for the Boonton gages above and below the reservoir. Rutger’s monthly climate tables show that 7.5 inches of snow fell on the town of Boonton during the month of January (Office of the NJ State Climatologist, 1994-2017b). During this flood, rain likely melted the snow cover. During the month of the flood, the Boonton reservoir was near capacity at about 7,500 MG (NJGS, 2013). The reservoir was unable to provide storage during the flood, resulting in a high peak discharge for the Rockaway River below Boonton gage. The Wanaque Reservoir was at a lower capacity of about at 27,408 MG during the month of January, and so the Wanaque gage downstream of the reservoir was not impacted by the flood.

The flood of September 1999 was caused by Tropical Storm Floyd (Martin and others, 2011). This flood demonstrates the possible differences in regional versus local precipitation, as the precipitation at Little Falls exceeded regional precipitation by almost 7 inches (Office of the NJ State Climatologist, 1994-2017a; Office of the NJ State Climatologist, 1994-2017b). Though the tributaries peaked earlier than the Passaic River at Little Falls, the Pompton River at Pompton Plains exceeded the discharge at Little Falls by over 2,000 cfs. The month prior to the flood, the Wanaque reservoir was at a lower capacity of 15,630 MG, resulting in a lack of change in discharge from the Wanaque River at Wanaque gage during peak flooding (NJGS, 2013; USGS, 2017).

Hurricane Irene was responsible for two floods in August and September. Extremely heavy rainfalls fell on already saturated ground from precipitation during the three weeks prior to the hurricane (Watson et al., 2014). This caused historic levels of flooding within the Passaic River Basin, and two major flooding events only a week apart. Many homes and residential areas were flooded and had to be evacuated. These heavy rainfalls caused both the Little Falls and Pompton gages to record their highest discharge peaks since 1903, at over 20,000 cfs (USGS, 2017). The Wanaque Reservoir's high water levels were a likely contributor to the Central Passaic and Pompton River's high discharges. 2011 was the wettest year on record, so it is not surprising that it produced the most floods, and the highest peak discharges of the past 100 years (Office of the NJ State Climatologist, 1994-2017a).

Differences between the precipitation prior to flooding and the discharge among floods occurring in the same season and year suggest that heavy precipitation is not the sole factor in major flooding. While heavy average monthly precipitation brought on by storms over the Climate Division 1 of New Jersey are the most common cause of major flooding within the Passaic River Basin, flooding factors vary among individual floods. Heavy local precipitation can cause major flooding for the Passaic, even if precipitation over Climate Division 1 is moderate. Precipitation for the months prior to flooding have the largest impact in the winter due to the melting of snow cover and ice (Paulson et al., 1991). Melting of snow and ice can cause major flooding even in the absence of heavy precipitation during the month of the flood. Continuous precipitation in warmer months can lead to oversaturation of the ground, resulting in major flooding during the next heavy rain. Development of urban areas and impervious surfaces would also have an effect on how precipitation would affect river discharge, but the magnitude of this effect is difficult to quantify.

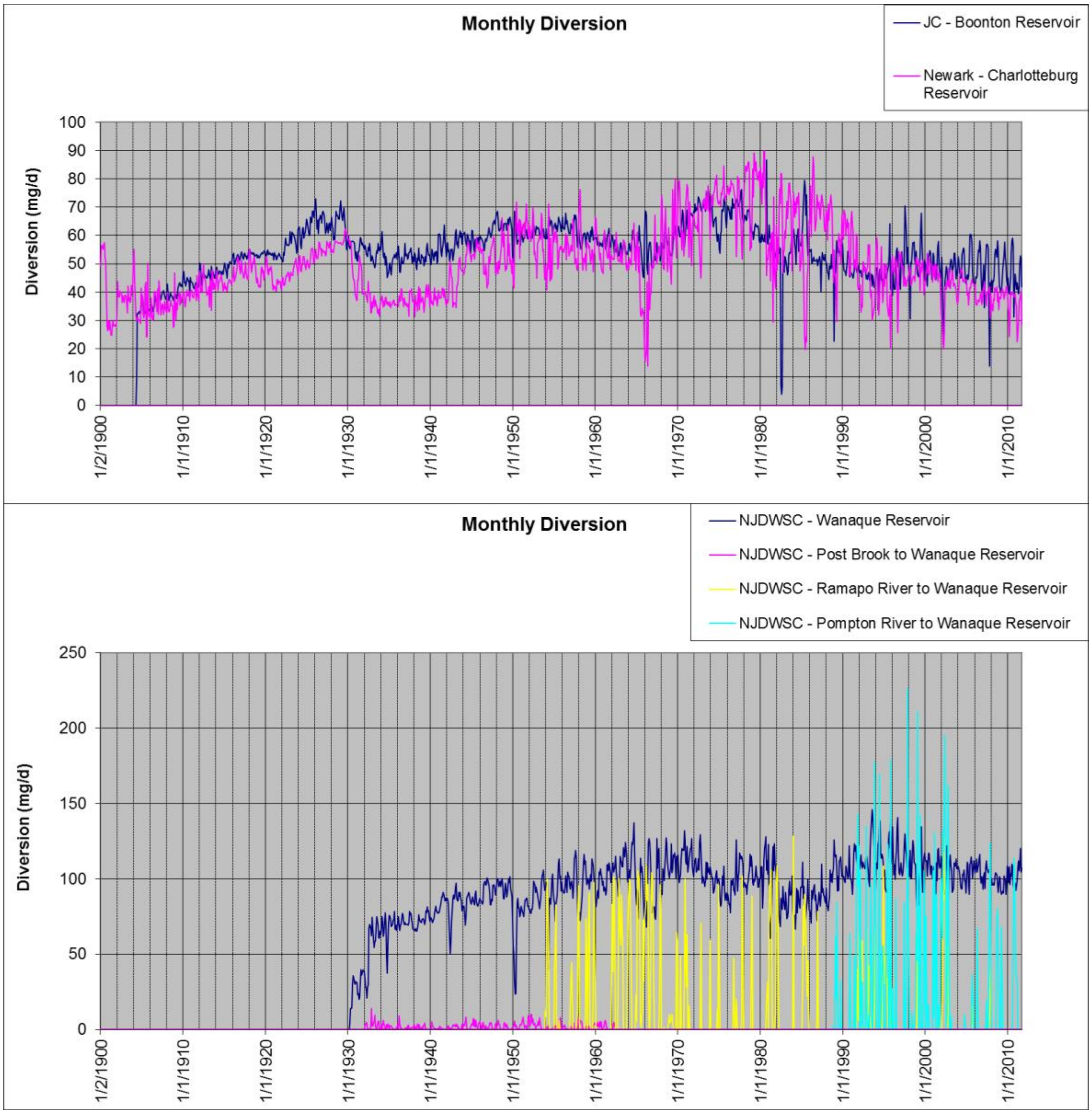


Fig. 30: Diversions by North Jersey District Water Supply Commission (top), Jersey City, and Newark (bottom) Data from New Jersey Geological Survey (NJGS) Digital Geodata Series (DGS) 2013 (NJGS, 2013)

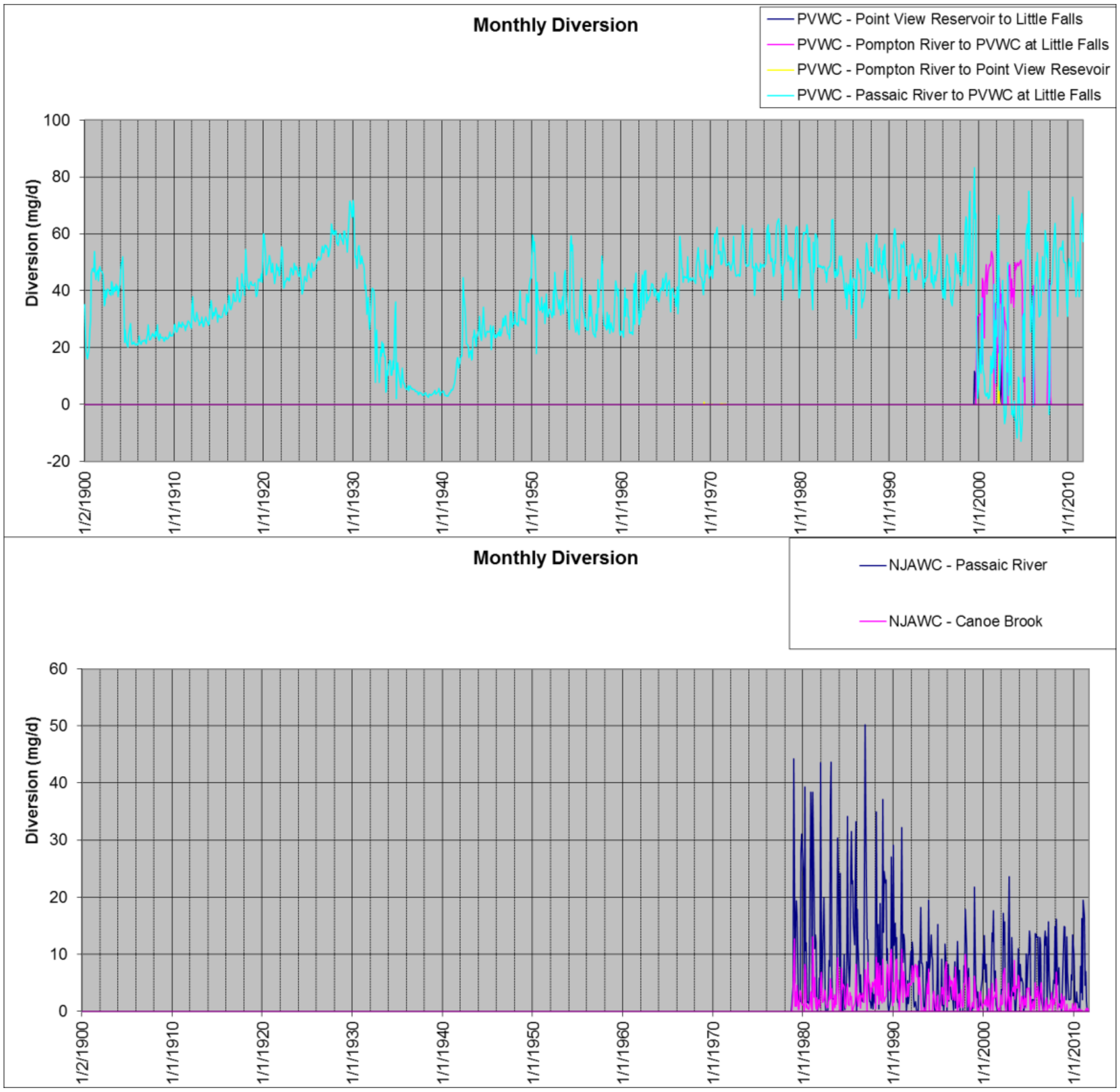


Fig. 31: Diversions for Company Passaic Valley Water Commission (top) and NJ American Water (bottom) Data from New Jersey Geological Survey (NJGS) Digital Geodata Series (DGS) 2013 (NJGS, 2013)

Anthropogenic Impacts Though heavy precipitation brought on by storms and hurricanes is the main cause of major flooding, there are anthropogenic factors affecting flooding as well. Because of the complexity of the basin and its long history of development and water supply demands, it is impossible for this study to quantify the total human impact upon flooding within the basin. However, the nature of human impacts is apparent in several data sets. For example, examination of peak flood discharge reveals the effect that reservoirs have on major flooding. The two gages located below reservoirs, the Wanaque River at Wanaque and the Rockaway River below Boonton Reservoir gages show the most variability in their discharge values (Fig. 15). While they usually follow peaks and declines with comparable values with the other tributary gages, there are also floods during which they have the lowest values that barely fluctuated during flooding. In the case of the Wanaque gage, low discharge values during major floods coincided with a deficit in reservoir storage (below 16 million gallons) during the month prior to flooding (NJGS, 2013; USGS, 2017). Higher discharge values with more prominent peaks occurred when the reservoir was close to capacity. This is true, to a lesser extent, of the gage below the smaller Boonton reservoir, although during most floods, the discharges from the gages both up and downstream of the reservoir are similar. This shows that reservoir capacity during flooding has a direct impact on discharge closely downstream, but that impact is dependent on the size of the reservoir. While it did not prevent major flooding, peak discharges were usually lower at the Little Falls gage when the reservoirs were at lower capacity. This shows that having reservoirs at lower capacities at least a month before a storm could decrease the intensity of flooding.

An increasing trend in annual runoff ratios in 8 of the 11 gages, despite reservoir storage and diversions, suggests influences from urbanization. Reservoir storage and streamflow diversions have also been shown to reduce runoff ratio values. Low runoff ratio values during drought years can also be attributed to increased diversion and storage of water for water supply. However, the rise in the average trend of the ratio despite reservoir construction and water diversions suggest other factors may also be contributing. Land development and urbanization are also factors in raising runoff. Besides forest clearing from colonial times and heavy industrialization in in the Lower Basin, suburban growth rose drastically after World War II, especially within the Central Basin. This lead to development encroaching on the Passaic's floodplains, and the loss of almost half the basin's wetlands from 1948-1978 (U.S. Environmental Protection Agency, 1994). An additional 4.5% of wetlands was also lost between 1985 and 2007 (Martin and others, 2011). This development not only increased impervious surfaces in the form of roads and parking lots and put residential and business buildings within the range of flooding, but reduced the basin's natural flood storage. This undoubtedly has an effect on runoff and flooding in the Lower and Central Basins. Studies within the Passaic River Basin show that an increase in impermeable surfaces does not have an immediate effect on increases in runoff (Martin and others, 2011). Only 17% of the land that increased in impervious surfaces between 2002 and 2007 had runoff increases over 50%. However, of those runoff increases, 35% had a 5-10 % increase on overall flow intensity, demonstrating that while land development over a short period of time may have little immediate impact on runoff, urbanization over time can lead to large cumulative increases in runoff volume and intensity (Martin and others, 2011).

It is also possible that land use changes had an effect on floodplain deposits seen in the tops of the cores as well. Although mercury concentrations suggest that deposits from the past 200 years are within the upper 20 cm of the cores, forests within the Great Piece Meadows were cleared for farming during colonial times (US Army Corps of Engineers, 1987). Reservoirs and dams built during the 1800s and 1900s also may retain channel sediments along the Passaic and its tributaries, preventing them from reaching areas downstream. Though not the main factor, restrictions of sediment transport and loss of forest cover may contribute to the lack of recent flood deposits containing high organic content or increased grain size within the central Passaic floodplain record.

CONCLUSIONS

The floodplain and flooding history of the Passaic River has been shaped by both natural and anthropogenic factors. The river's long, gently sloping course and its exit from the Central Basin through a gap in the Watchungs was shaped by its glacial history, while its wide, flat floodplains in the Central Basin underlain by silt and clay are a result of the formation of Glacial Lake Passaic (Salisbury and Kümmel, 1895; Stanford, 2007b). During storms, narrow, high gradient tributaries from the Highlands empty quickly into the low gradient Passaic River, and are retained there by the narrow outlet in the basalt bedrock at Little Falls, where infiltration is limited by underlying glacial lake clays (Hoffman, 1989; Hollister and Leighton, 1903). Analysis of peak discharges of the

Passaic and its tributaries during 27 major floods confirms this; showing flashy discharge of the tributaries a day or two before peak discharge and gradual streamflow reduction at Little Falls.

The Great Piece Meadows within the Central Passaic Basin contains several oxbow cutoffs, which contain a complete record of Holocene and modern floodplain sedimentation. The Oxbow2, RC2, and TZS2 cores all show a fining upward sequence from channel sand, to a mixture of silt and sand, and finally silt and silty-clay after the complete cut off oxbow and transition to a lake or wetland environment. Oxbow TZS, the oldest oxbow, was cut off from the main channel at around 9-9.5 ka B.P. Oxbows Oxbow2 and RC are thought to be contemporary, being separated from the main channel around 3.5-4 ka B.P. Core RC2 records about 8-9 major floods from about 3,600 B.P-3,200 B.P., as thin layers of wood or leaves, while core Oxbow2 records about 12-14 major floods as thin layers of organics or fine sand around the same time period. Increases in mercury concentrations in oxbow sediments identify the start of coal burning (~1835) at 6-7 depth in the TZS cores, 7-12 cm in the RC cores, and 17-18 cm in the Oxbow2 cores, confirming a complete record. Floods within the top 50 cm of each core lack identifiable deposits of organics or sand, showing a change in overbank deposits from the Holocene. While this change reflects a change in sedimentation in the oxbows as they shifted from lakes to wetlands, there may be other factors involved. Continued flattening of the river gradient by isostatic rebound, which is 0.4 m/km in this region, may be a factor but, climate change is a more likely cause (Stone et al., 1989). The estimated avulsion of oxbow TZS coincides with the start of a Holocene wet period in New Jersey around 9-10 k.a. B.P.(Getch et al., 2016) The cutoffs of oxbows RC and

Oxbow2 also occur during a shift from dry to wet conditions during the mid-Holocene transition around 5.5-3.5 ka B.P. This suggests regional climate changes may be a factor in the avulsions of the central Passaic River and possibly changes in floodplain sedimentation.

The residents of the Passaic River Basin have experienced major flooding recorded since colonial times, especially within the Lower and Central Basins (Brydon, 1974). The gage along the Passaic River at Little Falls has experienced at least major 37 floods since 1870, which have decreased in intensity due to reservoir construction, but appear to be increasing in frequency. Reservoirs have reduced flood discharges within the Passaic. At low storage capacity, the Boonton and Wanaque Reservoirs have shown to eliminate major flooding directly downstream and reduced major flooding of the Passaic during storms. However, they cannot prevent major flooding within the Passaic. Average annual precipitation is shown to be steadily rising over the past century, suggesting climate change could be a factor increased flooding. However, the rise in the average annual runoff ratio for eight stream gages suggests that urbanization also plays a role in flooding. Widespread destruction wetlands and forests has occurred over the past century, while urban and suburban construction has continued to grow along the Passaic's floodplain, increasing the amount of impervious surfaces. The Passaic Basin's geology make it naturally prone to flooding, and this study suggests there are anthropogenic factors that affect flooding as well.

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APPENDIX I

MERCURY CONCENTRATION ANALYSIS

Mercury Background

Atmospheric mercury is mobilized both by natural processes and human industrial and mining activities. During its natural biogeochemical cycle, mercury is emitted from natural and human processes, transported throughout the atmosphere, deposited onto land and waterbodies, and is either remobilized or ultimately sinks into deep ocean sediments (Selin, 2009). Mercury has an atmospheric residence time of about 0.5-2 years (Lin and Penhkonen, 1999; Lindqvist, 1985; Selin, 2009). Natural sources of mercury include volcanism, oceanic emissions, geothermal activity, mercury enriched rock emissions and geologic weathering. These natural emissions are estimated to contribute to 33% of the world's present mercury emissions (Martin and Nanos, 2016; Sexauer, 2003). Anthropogenic mercury emissions are released by the burning of fossil fuels and organics, gold mining and production, nonferrous and steel production, cement production, and mercury production (Selin, 2009). Preindustrial mercury levels determined from remote lake sediment cores provide estimates that present mercury deposition is 3-5 times greater than preindustrial deposition (Selin, 2009).The main form of mercury in the atmosphere by both anthropogenic and natural sources is elemental mercury (Hg0), but divalent mercury (HgII) and the compound dimethyl mercury (CH_3)2Hg may also be present (Lindqvist, 1985). Although there has been a wide range of measurements of atmospheric mercury, studies in the 1980s report that concentrations in remote areas ranged from 1-4 ng/m3 (Ferrara et al., 1982), while concentrations over industrial areas with coal powered plants are over 1000 ng/m3 (Lindqvist, 1985). Only about 10-20% of the emissions are deposited locally, with the higher percentage carried farther away to be deposited regionally and even globally (Lindqvist, 1985). Anthropogenic sources were estimated to have released 3,000,000 kg/year (3000 tons/year) in 1900, which tripled in the 1970s. Currently, anthropogenic sources are estimated to release a range from 2200-4000 tons/year (Kim and Kim, 1999; Martin and Nanos, 2016).

Most mercury is deposited by both wet and dry deposition as Hg(II), but 5-60% of deposited mercury is quickly revolatized into the atmosphere as Hg0. Water or snow surfaces have higher percentages of revolatized (Hintelmann et al., 2002; Selin, 2009). The remaining mercury continues to reside in the soil, where it is absorbed. However, mercury in solution may be carried away as runoff or methylated by microbes. Mercury also tends to favor living plants and organic matter within the soil, where is strongly binds to reduced sulfur groups. Plants receive mercury from the soil through their roots, while atmospheric mercury is deposited on leaves and is thought to be taken into the plant by gas exchange through the stomata (Lindberg et al., 1992; Selin, 2009). Other processes of the intake of mercury by plants involved the absorption and oxidation of Hg(0) (creating Hg (II)), or the absorption of Hg (II) or particulate mercury (Schroeder and Munthe, 1998). Through leaf fall, the mercury intake from plants is added to the soil. When deposited to sediments in a lake environment, Hg (II) is assumed to be relatively immobile because it strongly binds to organic material (Sander et al., 2007).

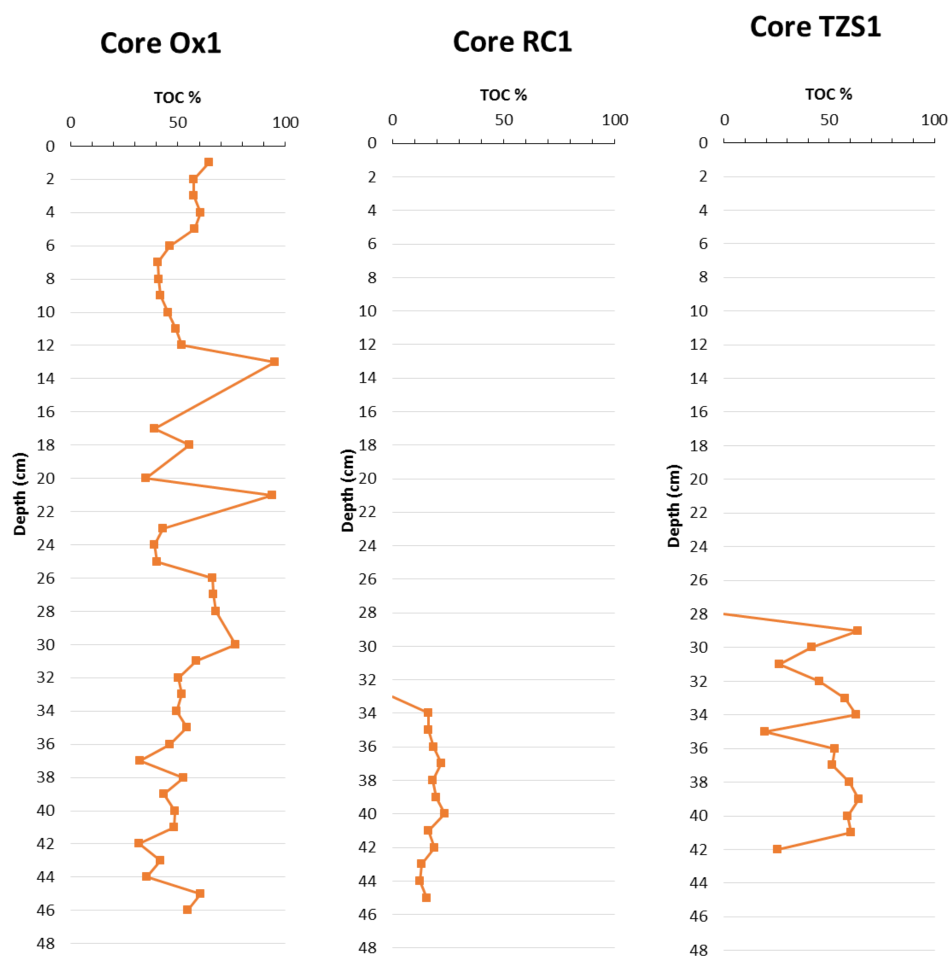
<u>Other Elemental Concentrations Used in Dating Methods</u>

Most mercury concentration profile studies are done with lake cores in conjunction with Pb210 dating. Pb210 is produced in soils either by the decay of Ra-226 associated with sediment particles, or by the decay of Rn-222 in the atmosphere, which can be dated (Kroenke et al., 2002). The half-life of Pb210 is 22.3 years, and ideally shows a peak at the surface of a sediment profile, with exponential decrease with depth. Although the cores taken from abandoned channel meanders that were once full oxbow lakes, they currently are situated in a wetland environment that frequently floods and retains water, but is not continuously submerged throughout the year. Because of this, the core cannot be expected to have had a steady sedimentation rate. Since Pb-210 is affected by particle mixing and does not record if sediments are retained within the drainage basin or wetland before being deposited, Pb-210 cannot be used as a dating tool for this study. Cs-137 associated with nuclear weapons testing in the 1950s is also used to date sediments, but not before 1950. Cs-137 peak levels are linked to maximum deposition from 1963-1964 (Kroenke et al., 2002).

Total Organic Carbon

Core OX 1A was also measured for total organic carbon by weighing its samples before and after being processed by the Hydra 2C (Appendix 1, Fig. 1). Since mercury binds strongly to organics, high organic compositions may contribute to elevated mercury concentrations. Samples with the highest percentage of organics (over 90%) were found at depths of 13 and 17 cm within the core, but do not correlate with particularly high mercury concentrations. Total organic carbon at depths of 28-40 cm ranged from 31-

77%. At depths of 29-42 centimeters of the core TZS 1A, total organic carbon ranged from 19-63%. At depths of 29-40 cm on core RC1A, total organic carbon ranged from 12-23%. Profiles of other mercury studies (Engstrom et al., 2007; Sander et al., 2007) also show a general trend of mercury concentrations gradually increasing after the onset of coal powered industry, but peaking higher up within the core. The same trends could be expected to be seen in the cores from the Great Piece Meadows.



Appendix I, Fig. 1: Total organic carbon for cores OX1, RC1, and TZS1 as analyzed by the Hydra IIC mercury analyzer

APPENDIX II

FLOOD DEPOSITS

Flood Deposit Identification

Prior to this study, identifying features of paleoflood deposits within the Central Passaic River floodplain was unknown. Analysis of the oxbow cores suggest that there has been a change in major flood deposits within the oxbows. It is impossible to determine deposits of individual recent major floods from the Passaic oxbow cores since mercury analysis suggests that the record of the past 200 years occurs within the top 20 cm of the core, and the layer of clayey-silt at the top of the RC and Oxbow cores is too thick to be associated with any one particular flood. However, the tops of all three cores consist of silt or clayey-silt, showing that fine grained sediments are associated with the recorded major floods starting in the 1800s. Thin layers of coarser sediment and dense organics appear to be indicative of ancient major floods. Though correlating ancient flood deposits to modern flood discharge values is not feasible for this study, the Passaic achieves bankfull flow at 4,000 cfs at Little Falls and 2,000 cfs at Pine Brook, giving a minimum reference for ancient flood discharges (Vermeule, 1894). As a reference for river channel sediments, a sample of sand was taken just off the river bank of the current channel of the Passaic River, less than a mile upstream of the core TZS site was determined, by weight, to be a fine-medium sand. This sample was taken from off the banks of the Passaic, and may not be representative of all channel sediments. It can be assumed, though, that medium-fine sand layers within the cores are indicative of channel or reactivated meander cutoff sediments.

Previous studies and papers of flood and slackwater deposits corroborate the identification of flood deposits within the Central Passaic River Basin. Slackwater deposits, which consist mostly of sand and silt, quickly accumulate from suspension during major floods, and therefore can be used in identifying paleofloods (Baker, 1987). Silt-clay and organic drapes within the stratigraphic record are detritus from the last flood inundation. Organic layers are litter that has accumulated on the surface of the floodplain in between flood events. Sharp variations in vertical grain size indicate sediment emplacement by individual flood events (Baker, 1987). Depositional layers of seeds, leaves, and small twigs are likely seasonal ground litter, and will have a radiocarbon age within one year of the flood that transported it (Baker, 1987; Baker et al., 1985)

Flood deposits from recent studies in areas with similar geology were also taken into account. In a study examining the effects of flooding from hurricane Irene on the sediment deposits within oxbow lakes of the Connecticut River, a notable difference in the sediment record was observed after the flood (Yellen et al., 2014). A new layer of organic poor clayey-silt was deposited in oxbow lakes after the flood, which was finer grained and lower in organics than the underlying sediment layer. In the core from Chapman Pond in the Connecticut River Basin, organics were about 50% lower in the medium silt layer from Hurricane Irene, than in the very fine sand layer from the previous Spring freshet. Sediment yield during Hurricane Irene within the Deerfield River Basin (a tributary of the Connecticut River) was estimated to be equivalent of 10-40 years of sedimentation for average discharge values, and five times the amount for average sedimentation of the Connecticut River watershed (Yellen et al., 2014). Although an increase in coarser grained sediment is usually indicative of flooding, the increase in fine grained sediment in the Connecticut River Basin from Hurricane Irene is likely explained by the extensive supply of fine grained material within the eroded till and glaciolacustrine deposits at higher elevations within the watershed, and are a result of headwater erosion. The lower lying areas of the floodplain lack evidence of the hurricane within their sedimentary record (Yellen et al., 2014). The Connecticut River, like the Passaic River, also lies within a postglacial watershed. Low lying streams, such as the Connecticut and Passaic Rivers, often feature sediments washed in from high elevation tributaries by major events. While coarse glacial till and bedrock is dominant along the courses of the tributaries, lacustrine and wetland silt and clay deposits are also present, and may contribute to fine grained flooding deposits (Stanford, 1993).

APPENDIX III

GRAIN SIZE

Table 1:

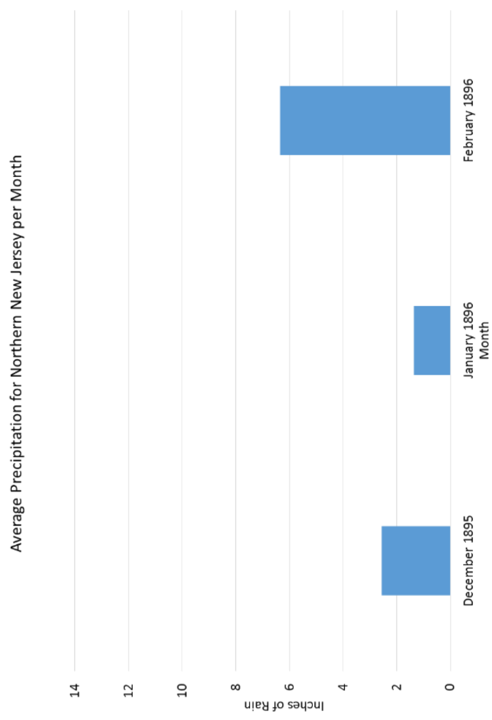
Core	Sample Name	Depth (cm)	% <2µm	% 2-50µm	% 50-100µm	% 100-250µm	% 250-500µm	% 500-1000µm	% 1000-2000µm
TZS2	TZS2A-Sand2	21	15.07	79.17	3.64	1.52	0.54	0.06	
TZS2	TZS2A-Sand3	35-36	10.19	83.63	4.87	1.08	0.15	0.07	0.02
TZS2	TZS2B-Sand1	58	14.69	81.28	2.44	0.93	0.04	0.29	0.28
TZS2	TZS2B-Sand2	65	9.37	75.29	9.59	5.6	0.15	0	0
TZS2	TZS2B-Sand3	153.5	1.34	7.49	6.58	42.65	32.84	9.1	0
TZS2	TZS2B-Sand4	91	13.71	77.45	5.42	2.98	0.43	0	0
TZS2	TZS2C-Sand1	104-105	13.12	74.54	8.3	3.95	0.09	0	0
TZS2	TZS2C-Sand2	109-110	13.36	73.36	10.45	3.62	0.26	0.04	0
TZS2	TZS2C-Sand3	120-121	6.57	44	22.5	26.01	0.92	0	0
TZS2	TZS2D-Sand1	138	3.38	20.66	19.47	50.14	6.34	0	0
TZS2	TZS2D-Sand2	148	1.17	6.35	5.69	41.64	33.93	10.52	0.7
TZS2	TZS2D-Sand3	153.5	1.34	7.49	6.58	42.65	32.84	9.1	0
TZS2	TZS2D-Sand4	161	1.68	10.81	5.95	31.91	39.59	10.06	0
OX2	OX2A-Sand1	4.5	13.72	74.16	9.85	1.48	0.11	0	0.29
OX2	OX2A-Sand2	39	10.62	78.94	7.42	2.81	0.21	0	0
OX2	OX2A-Sand3	13	19.15	76.77	2.87	0.35	0.33	0.53	0.01
OX2	OX2A-Sand4	20	20.37	75.79	2.67	0.24	0.47	0.44	0.02
OX2	OX2A-Sand5	34	11.29	79.62	6.28	2.61	0.2	0	0
OX2	OX2A-Sand6	46	9.5	74.86	7.26	3.57	0.45	0.49	2.46
OX2	OX2B-Sand1	56.26	10.09	75.47	9.37	4.89	0.18	0	0
OX2	OX2B-Sand2	65.75	10.41	82.69	5.93	0.97	0	0	0
OX2	OX2B-Sand3	67.25	7.74	63.66	12.67	14.76	1.16	0	0
OX2	OX2B-Sand4	70.5	7.75	61.13	16.24	14.52	0.37	0	0
OX2	OX2B-Sand5	84.75	10.27	71.77	10.54	7.19	0.23	0	0
OX2	OX2B-Sand6	91	11.24	57.78	14.54	14.18	0.6	0.37	0.93
OX2	OX2C-Sand1	98	12.94	73.93	8.49	4.23	0.41	0	0
OX2	OX2C-Sand1(2)	97.75	11.06	70.36	10.15	7.69	0.74	0	0
OX2	OX2C-Sand2	102	7.9	40.07	21.75	28.85	1.44	0	0
OX2	OX2C-Sand2(2)	103.5	5.34	27.48	18.83	43.59	4.75	0	0
OX2	OX2C-Sand3	108.5	3.42	16.09	21.96	52.68	5.85	0	0
OX2	OX2C-Sand3(2)	105.5	2.9	15.34	22.32	53.64	5.8	0	0
OX2	OX2C-Sand4	112.25	10.9	63.66	11.68	7.93	1.09	0.95	2.74
OX2	OX2C-Sand4(2)	112	7.56	37.14	20.1	33.06	2.13	0	0
OX2	OX2C-Sand5	116.25	10.87	50.13	13.75	18.18	0.78	0.66	3.74
OX2	OX2C-Sand5(2)	116	3.47	13.92	14.17	59.9	8.54	0	0
OX2	OX2C-Sand5(2)	133	4.92	19.94	14.22	54.31	6.6	0	0
OX2	OX2C-Sand6(2)	133.5	8.01	35.4	18.46	35.9	2.23	0	0
OX2	OX2C-Sand7	149.5	6.35	21.9	14.22	51.78	5.78	0	0
OX2	OX2C-Sand8	158	8.32	43.65	17.3	28.85	1.87	0	0
OX2	OX2C-Sand8(2)	158.5	7.71	40.15	16.67	32.06	3.41	0	0
OX2	OX2C-Sand9	171	9.94	43.95	8.09	13.99	12.39	2.39	6
OX2	OX2C-Sand9(2)	171.25	3.83	16.58	5.26	50.19	24.13	0	0
OX2	OX2D-Sand1	185	9.87	79.37	6.01	2.06	0.1	0.5	1.4
OX2	OX2D-Sand2	188.25	8.63	64.82	13.17	12.88	0.49	0	0
OX2	OX2D-Sand3	190	5.41	28.47	18.34	40.76	4.26	0	1.23
OX2	OX2D-Sand4	191	6.4	33.09	16.84	39.19	4.48	0	0
OX2	OX2D-Sand5	194	7.59	57.08	16.01	18.48	0.85	0	0
OX2	OX2D-Sand6	197.5	3.55	18.26	7.29	43.28	27.37	0.24	0
OX2	OX2D-Sand7	201.75	3.76	14.31	5.49	44.59	31.47	0.38	0
OX2	OX2D-Sand8	206.25	2.13	9.14	2.5	42.35	42.75	1.12	0

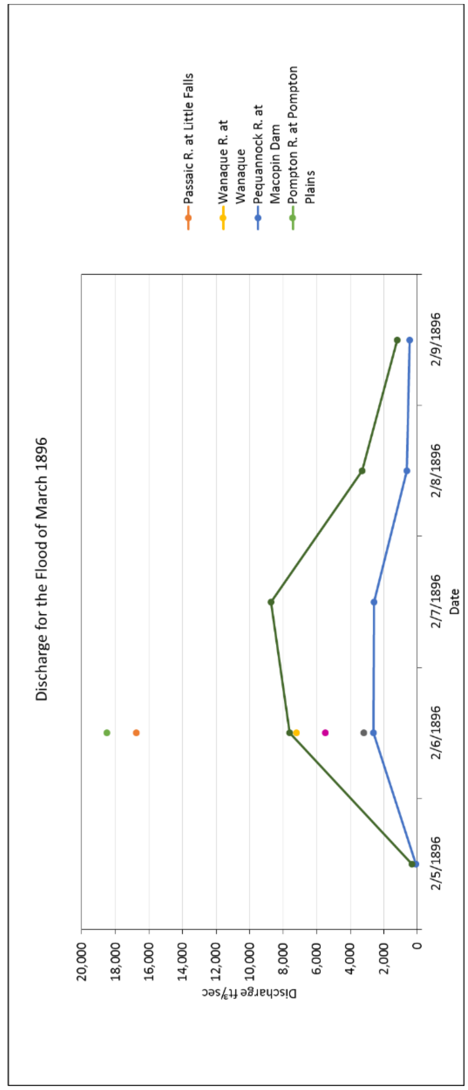
RC2	RC2A-Sand1	17	13.62	74.25	8.47	3.49	0.17	0	0
RC2	RC2A-Sand2	27	11.16	79.06	6.49	2.87	0.05	0.34	0.03
RC2	RC2A-Sand3	31-32	13.14	78.27	7.22	0.89	0.41	0.06	0
RC2	RC2B-Sand1	39-40	12.82	75.17	8.5	2.95	0.53	0.03	0
RC2	RC2B-Sand2	42.5	11.93	75.99	7.88	2.38	0	0.07	1.07
RC2	RC2B-Sand3	82	10.92	81.7	5.36	1.89	0.13	0	0
RC2	RC2B-Sand4	122	9.94	77.08	9.42	3.16	0.35	0.04	0
RC2	RC2C-Sand1	137	11.13	77.24	7.56	3.72	0.34	0.01	0
RC2	RC2C-Sand2	139	11.76	75.48	5.61	1.34	0	0.68	3.37
RC2	RC2C-Sand3	162	10.85	73.59	12.15	2.85	0.45	0.1	0
RC2	RC2C-Sand4	173	7.42	49.53	21.37	19.38	2.3	0	0
RC2	RC2C-Sand5	183	4.56	37.9	29.91	26.49	1.14	0	0
RC2	RC2C-Sand6	191	2.66	14.92	32.19	47.39	2.84	0	0
RC2	RC2C-Sand8	202	38.76	61.16	0.09	0	0	0	0
RC2	RC2C-Sand9	208	36	64	0	0	0	0	0

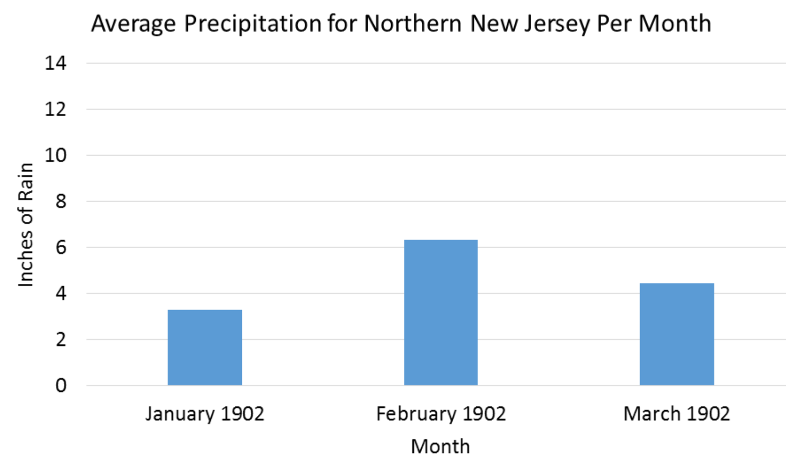
APPENDIX IV

FLOOD HYDROGRAPHS

Discharge data taken from: (Hollister and Leighton, 1903; Leighton, 1904; Salisbury and Knapp, 1897; Smock, 1891; USGS, 2017; Vermeule, 1894)


Average Precipitation for Northern New Jersey per Month
Inches of Rain
0
2
4
6
8
10
12
14
December 1895
January 1896
February 1896
Month


Discharge for the Flood of March 1896
Discharge ft^3/sec
0
2,000
4,000
6,000
8,000
10,000
12,000
14,000
16,000
18,000
20,000
2/5/1896
2/6/1896
2/7/1896
2/8/1896
2/9/1896
Date
Passaic R. at Little Falls
Wanaque R. at Wanaque
Pequannock R. at Macopin Dam
Pompton R. at Pompton Plains



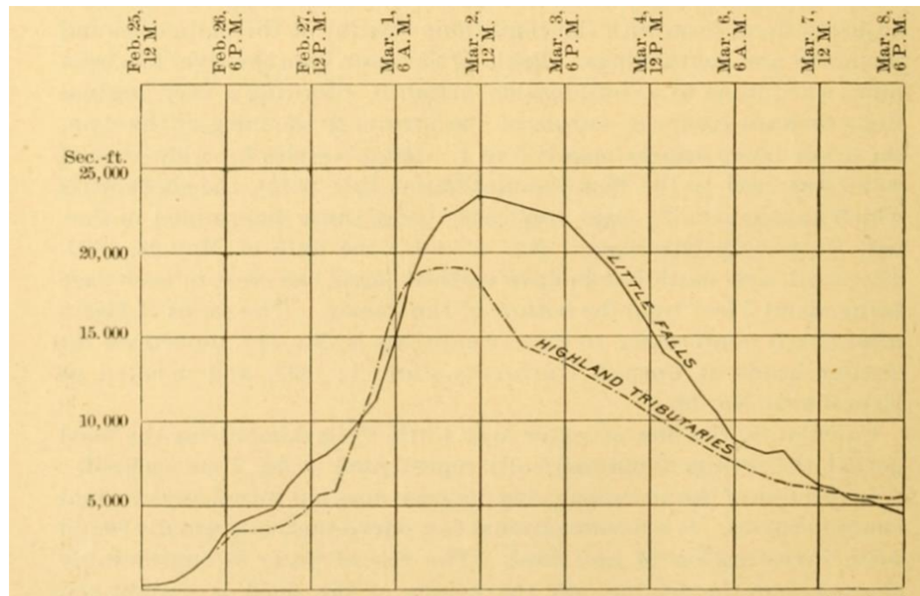
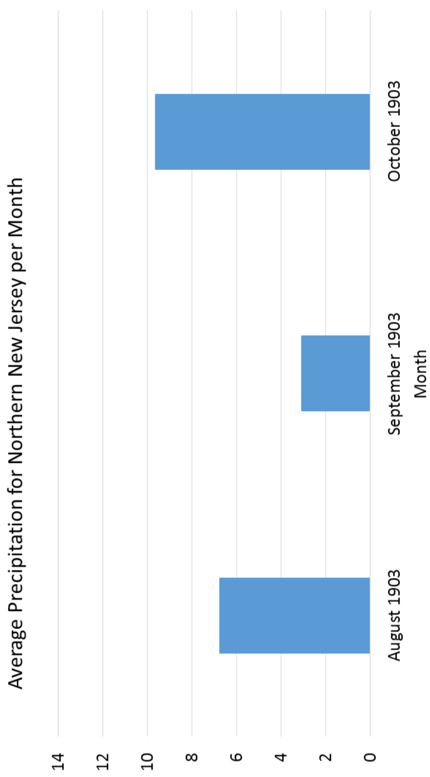
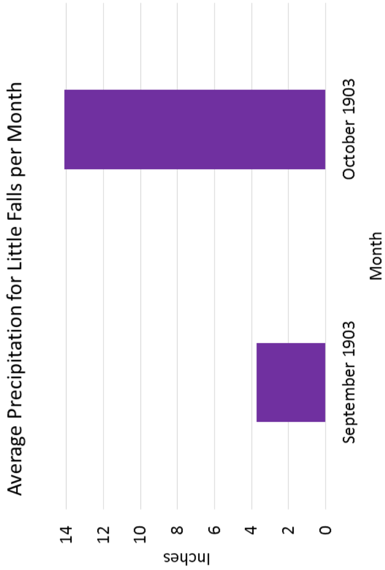
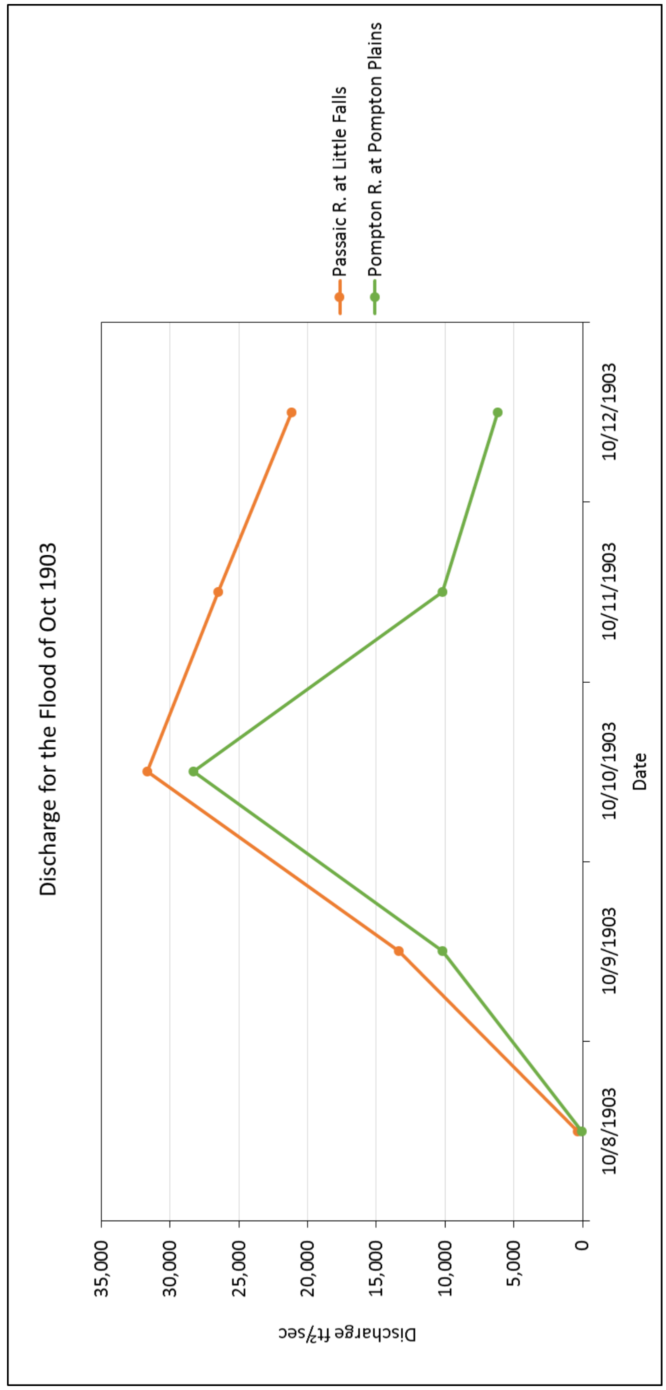
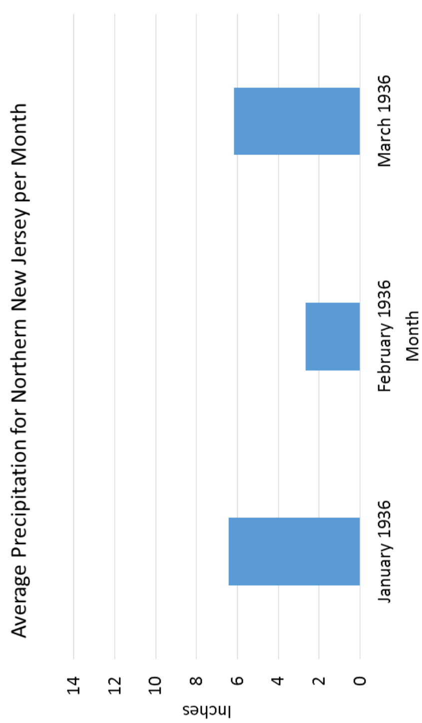


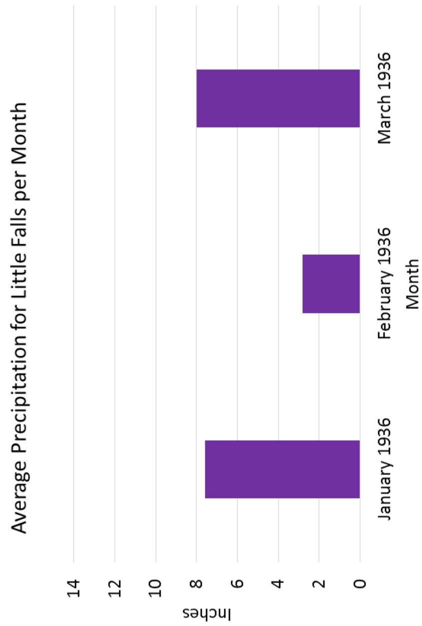
FIG. 8.—Comparative diagram of the flow from Highland tributaries into the central basin of the Passaic and the flow from Beattie's dam at Little Falls.

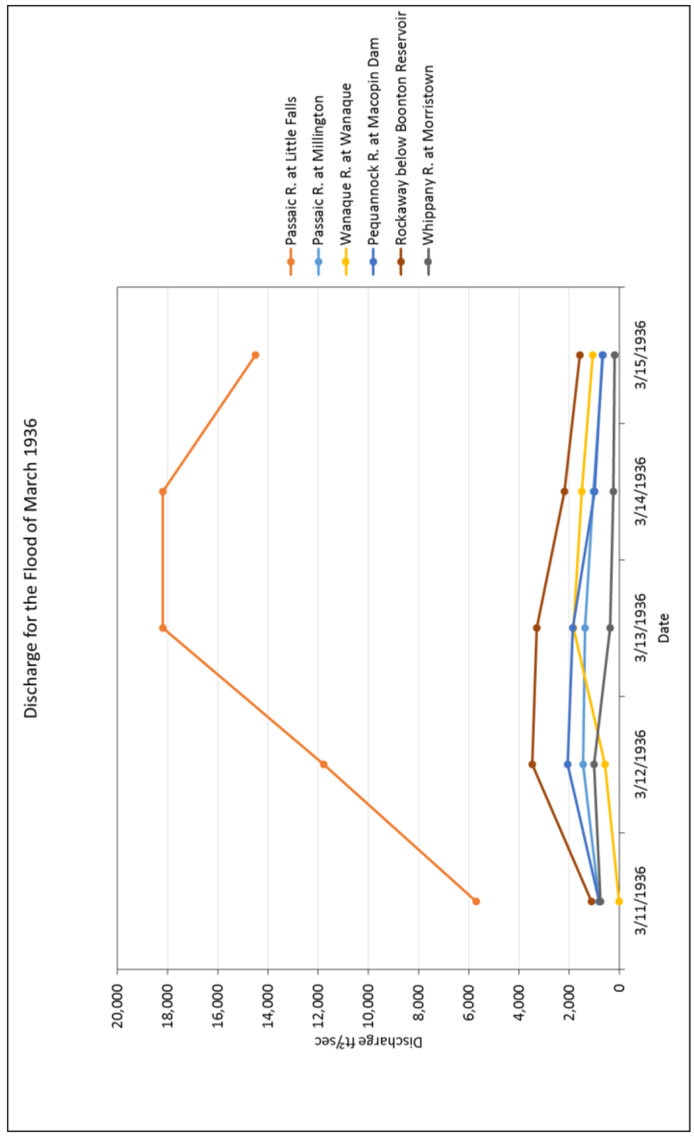

Average Precipitation for Northern New Jersey per Month
Inches
14
12
10
8
6
4
2
0
August 1903
September 1903
October 1903
Month

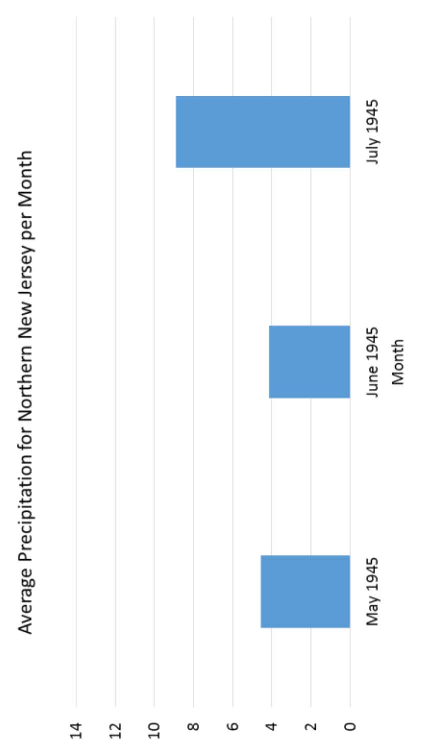

Average Precipitation for Little Falls per Month
Inches
14
12
10
8
6
4
2
0
September 1903
October 1903
Month

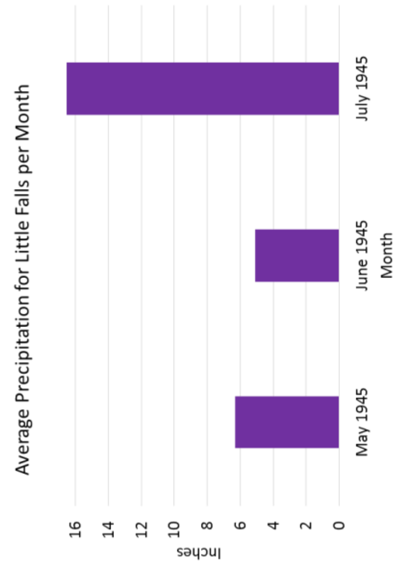

Discharge for the Flood of Oct 1903
Discharge ft3/sec
35,000
30,000
25,000
20,000
15,000
10,000
5,000
0
10/8/1903
10/9/1903
10/10/1903
10/11/1903
10/12/1903
Date
Passaic R. at Little Falls
Pompton R. at Pompton Plains

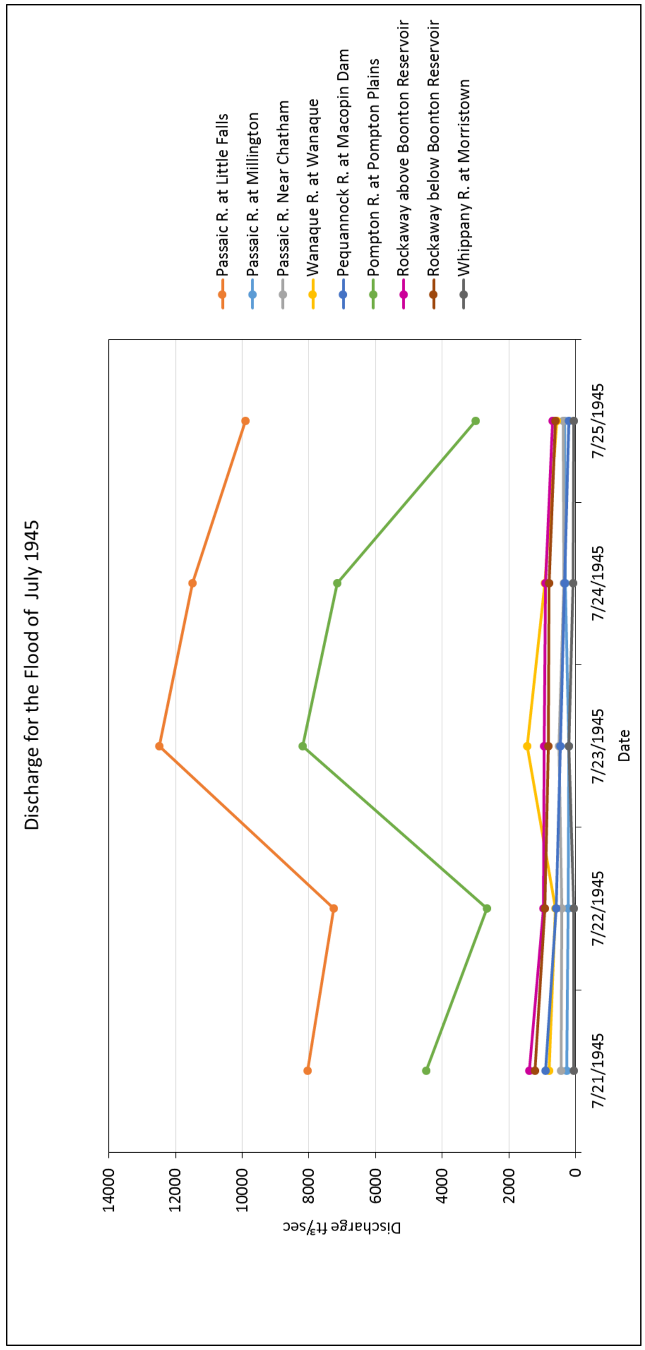

Average Precipitation for Northern New Jersey per Month
14
12
10
8
6
4
2
0
Inches
January 1936
February 1936
March 1936
Month

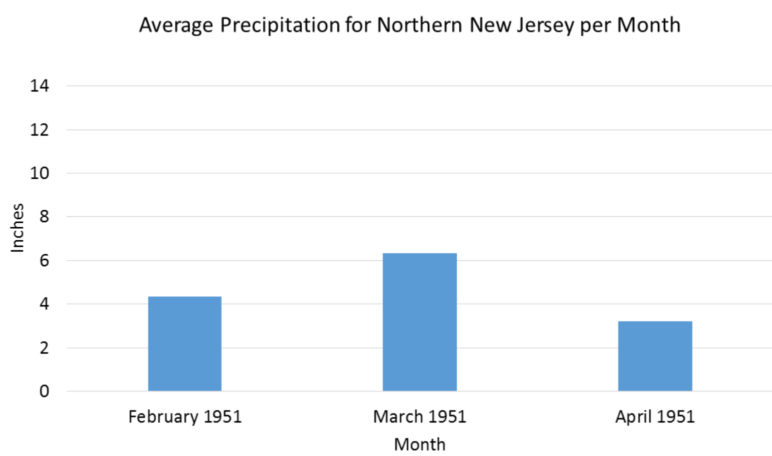

Average Precipitation for Little Falls per Month
14
12
10
8
6
4
2
0
Inches
January 1936
February 1936
March 1936
Month

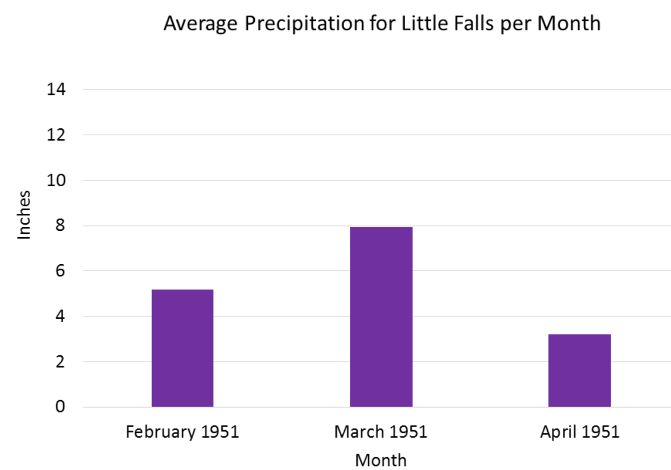

Discharge for the Flood of March 1936
20,000
18,000
16,000
14,000
12,000
10,000
8,000
6,000
4,000
2,000
0
Discharge ft³/sec
3/11/1936
3/12/1936
3/13/1936
3/14/1936
3/15/1936
Date
Passaic R. at Little Falls
Passaic R. at Millington
Wanaque R. at Wanaque
Pequannock R. at Macopin Dam
Rockaway below Boonton Reservoir
Whippany R. at Morristown

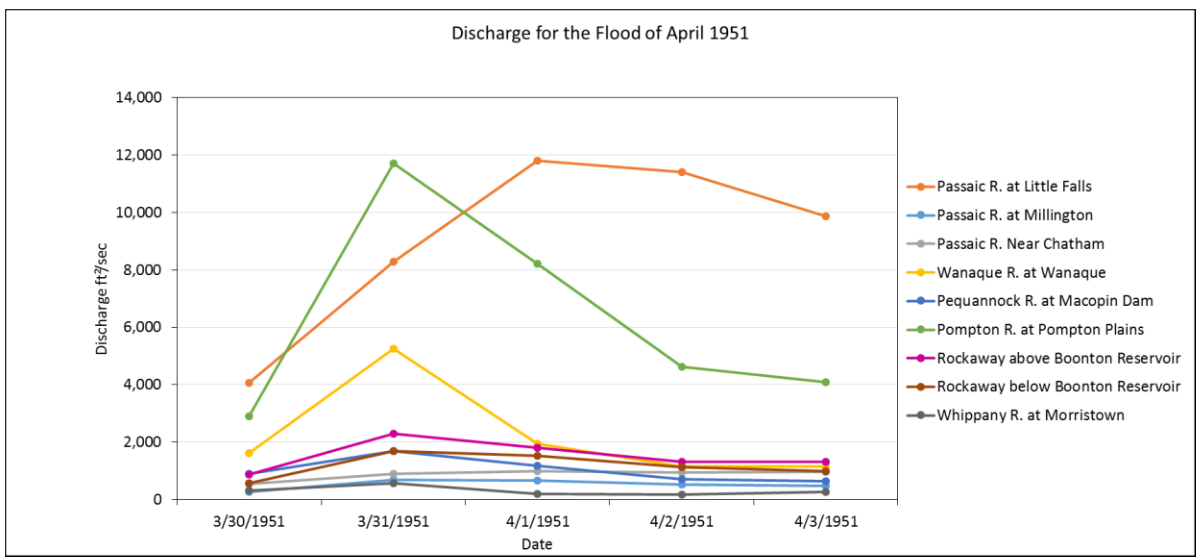

Average Precipitation for Northern New Jersey per Month
Inches
0
2
4
6
8
10
12
14
May 1945
June 1945
July 1945
Month

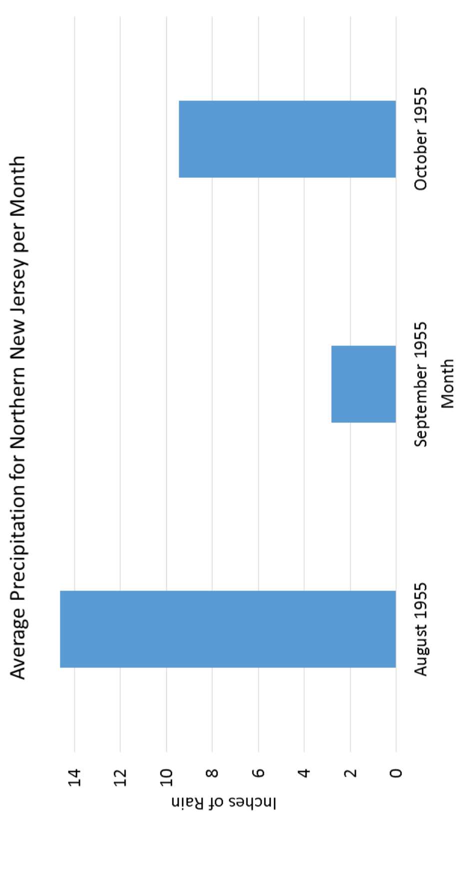

Average Precipitation for Little Falls per Month
Inches
0
2
4
6
8
10
12
14
16
May 1945
June 1945
July 1945
Month

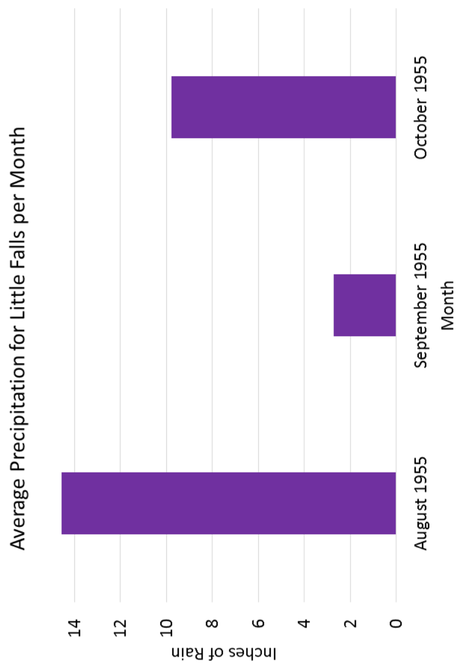

Discharge for the Flood of July 1945
Discharge ft³/sec
0
2000
4000
6000
8000
10000
12000
14000
7/21/1945
7/22/1945
7/23/1945
7/24/1945
7/25/1945
Date
Passaic R. at Little Falls
Passaic R. at Millington
Passaic R. Near Chatham
Wanaque R. at Wanaque
Pequannock R. at Macopin Dam
Pompton R. at Pompton Plains
Rockaway above Boonton Reservoir
Rockaway below Boonton Reservoir
Whippany R. at Morristown

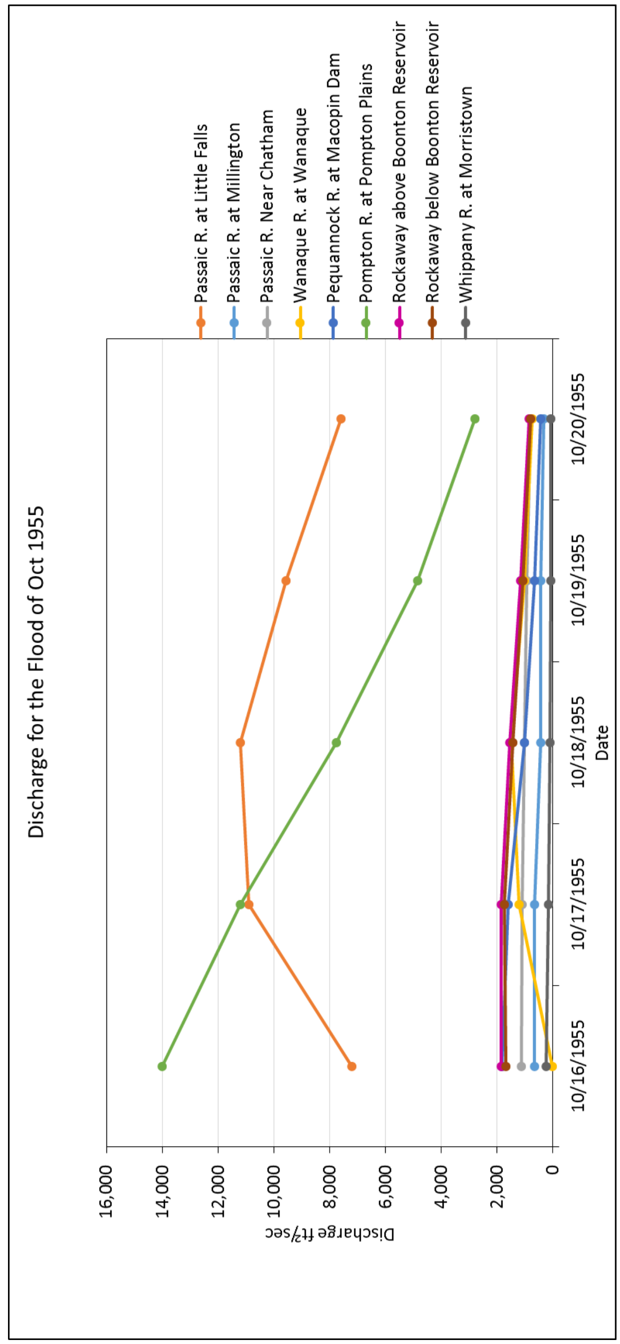

Average Precipitation for Northern New Jersey per Month
Inches
14
12
10
8
6
4
2
0
February 1951
March 1951
April 1951
Month

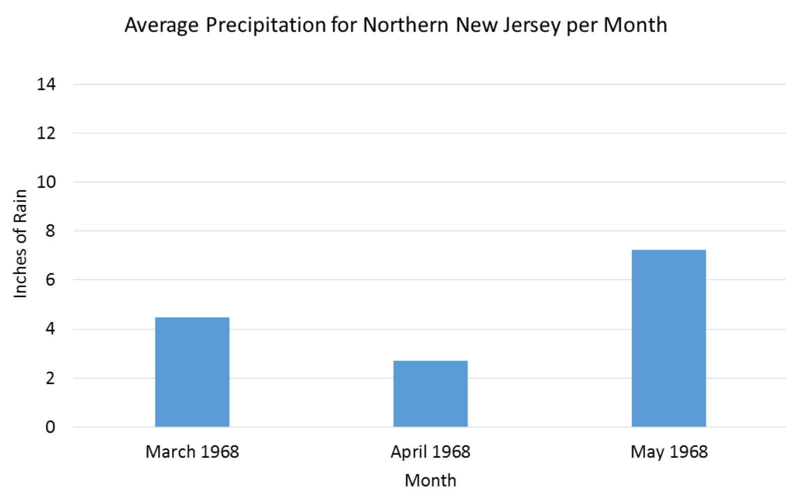

Average Precipitation for Little Falls per Month
Inches
14
12
10
8
6
4
2
0
February 1951
March 1951
April 1951
Month

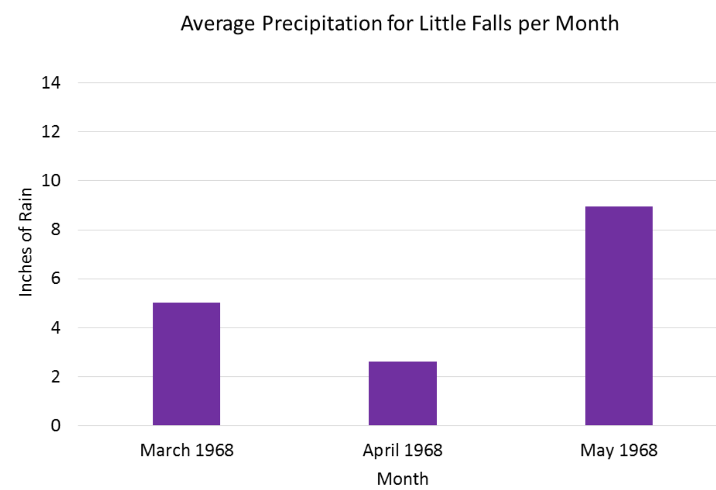

Discharge for the Flood of April 1951
Discharge ft³/sec
14,000
12,000
10,000
8,000
6,000
4,000
2,000
0
3/30/1951
3/31/1951
4/1/1951
4/2/1951
4/3/1951
Date
Passaic R. at Little Falls
Passaic R. at Millington
Passaic R. Near Chatham
Wanaque R. at Wanaque
Pequannock R. at Macopin Dam
Pompton R. at Pompton Plains
Rockaway above Boonton Reservoir
Rockaway below Boonton Reservoir
Whippany R. at Morristown

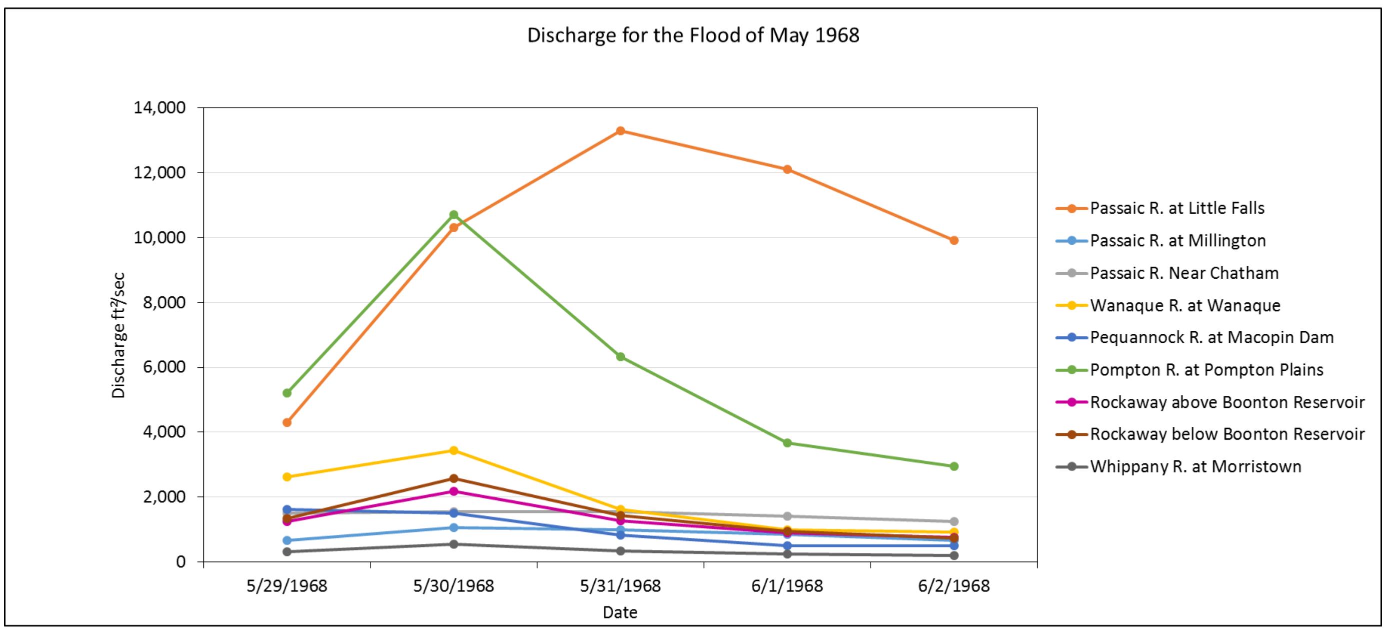

Average Precipitation for Northern New Jersey per Month
Inches of Rain
14
12
10
8
6
4
2
0
August 1955
September 1955
October 1955
Month

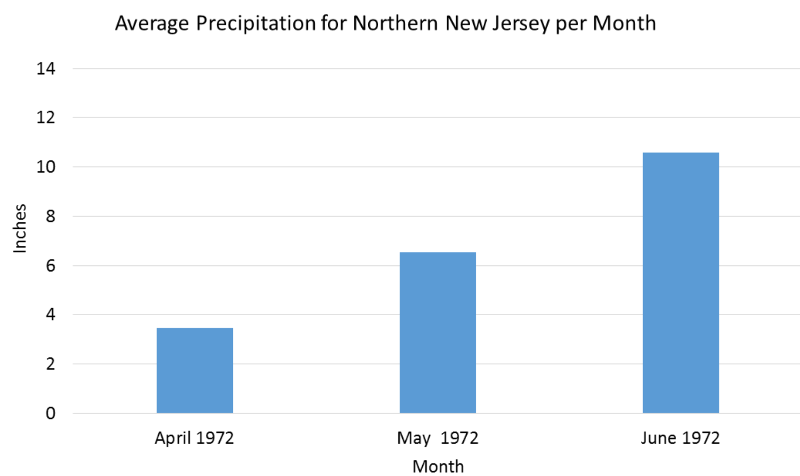

Average Precipitation for Little Falls per Month
Inches of Rain
14
12
10
8
6
4
2
0
August 1955
September 1955
October 1955
Month

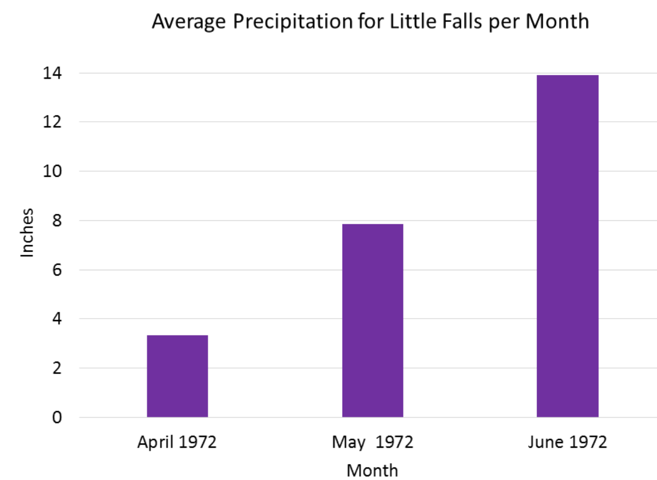

Discharge for the Flood of Oct 1955
Discharge ft³/sec
16,000
14,000
12,000
10,000
8,000
6,000
4,000
2,000
0
10/16/1955
10/17/1955
10/18/1955
10/19/1955
10/20/1955
Date
Passaic R. at Little Falls
Passaic R. at Millington
Passaic R. Near Chatham
Wanaque R. at Wanaque
Pequannock R. at Macopin Dam
Pompton R. at Pompton Plains
Rockaway above Boonton Reservoir
Rockaway below Boonton Reservoir
Whippany R. at Morristown

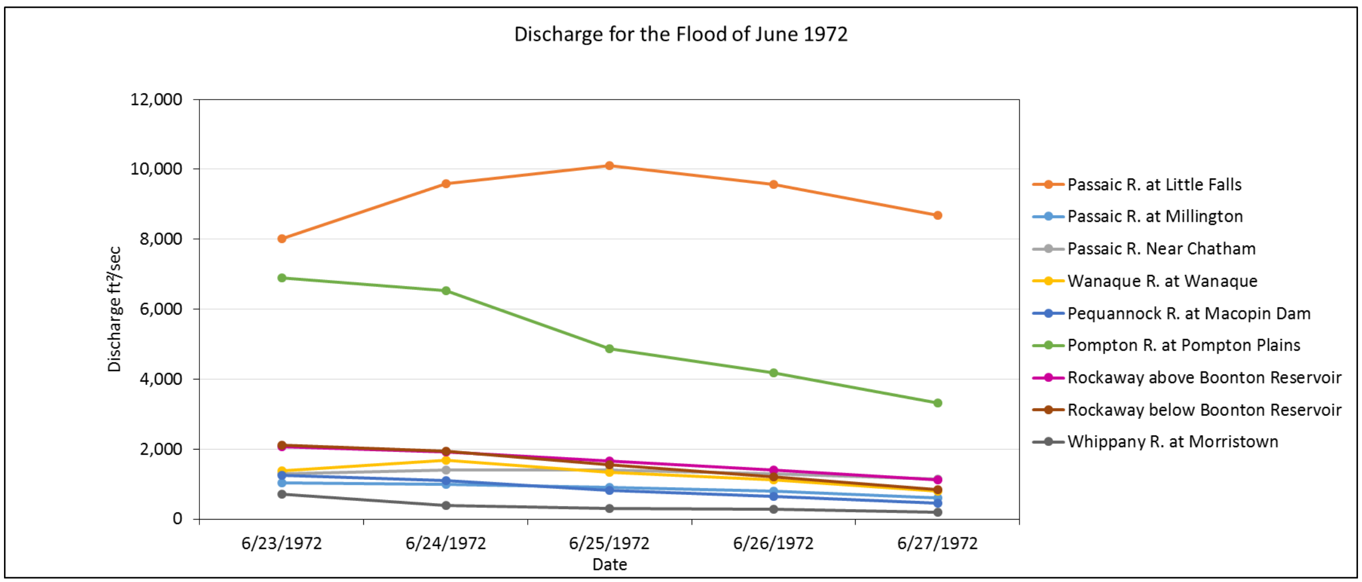

Average Precipitation for Northern New Jersey per Month
Inches of Rain
14
12
10
8
6
4
2
0
March 1968
April 1968
May 1968
Month

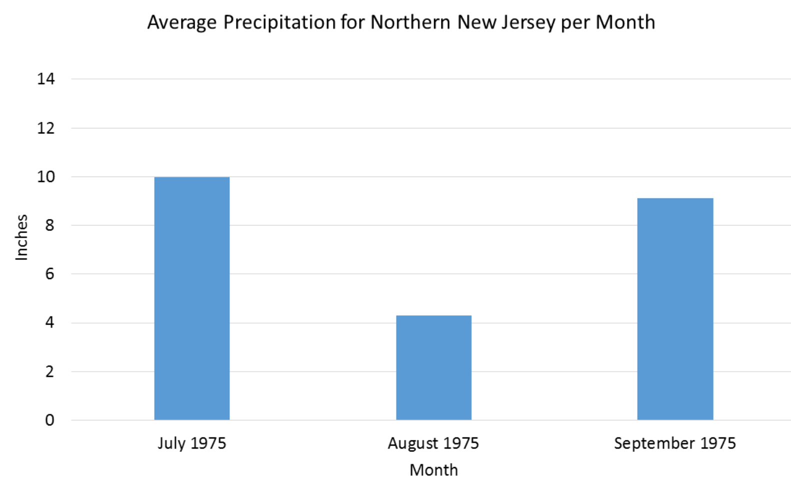

Average Precipitation for Little Falls per Month
Inches of Rain
14
12
10
8
6
4
2
0
March 1968
April 1968
May 1968
Month

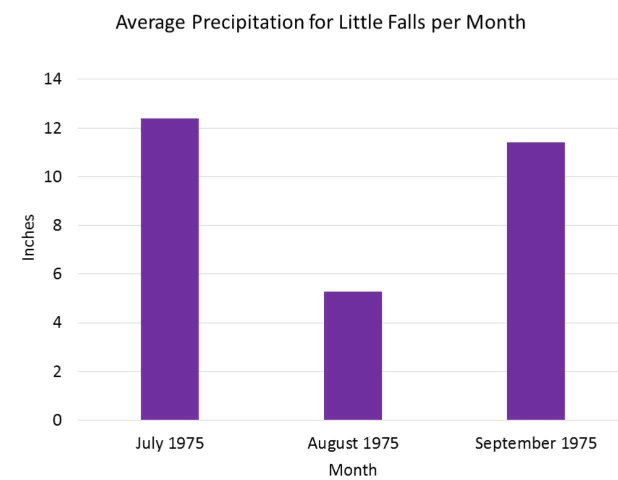

Discharge for the Flood of May 1968
Discharge ft³/sec
14,000
12,000
10,000
8,000
6,000
4,000
2,000
0
5/29/1968
5/30/1968
5/31/1968
6/1/1968
6/2/1968
Date
Passaic R. at Little Falls
Passaic R. at Millington
Passaic R. Near Chatham
Wanaque R. at Wanaque
Pequannock R. at Macopin Dam
Pompton R. at Pompton Plains
Rockaway above Boonton Reservoir
Rockaway below Boonton Reservoir
Whippany R. at Morristown

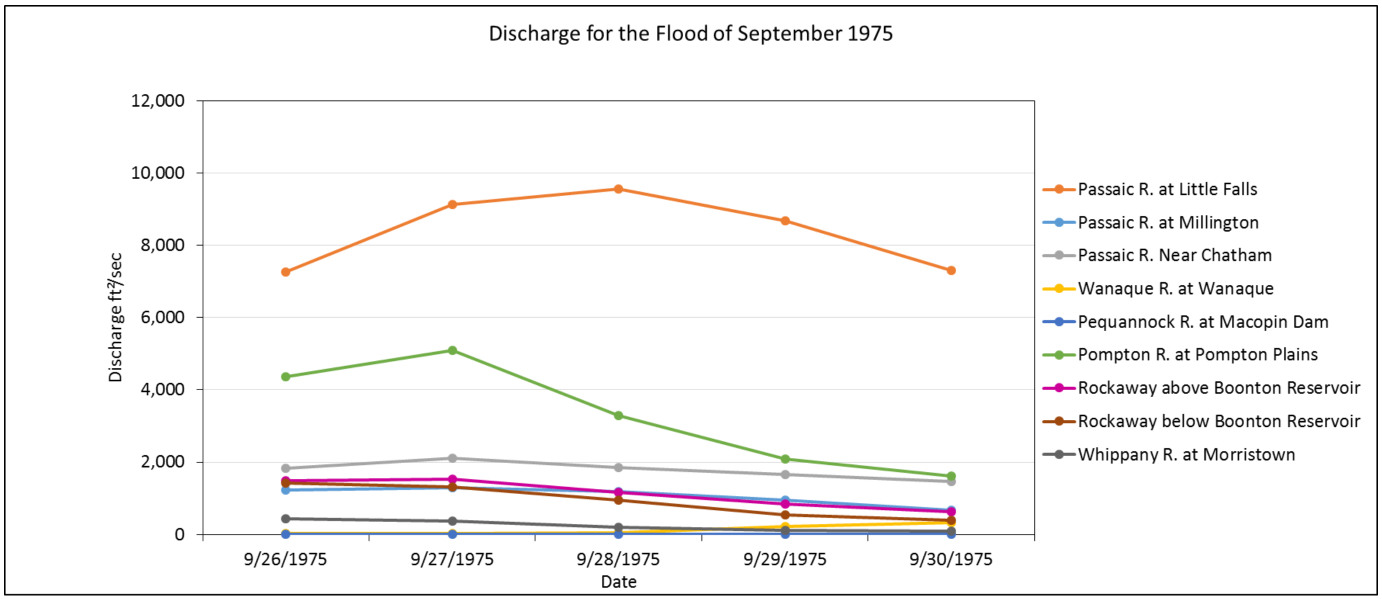

Average Precipitation for Northern New Jersey per Month
Inches
14
12
10
8
6
4
2
0
April 1972
May 1972
June 1972
Month

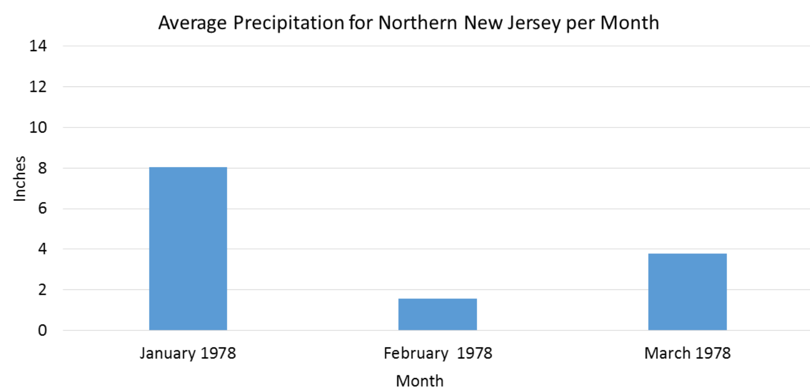

Average Precipitation for Little Falls per Month
Inches
14
12
10
8
6
4
2
0
April 1972
May 1972
June 1972
Month


Discharge for the Flood of June 1972
Discharge ft³/sec
12,000
10,000
8,000
6,000
4,000
2,000
0
6/23/1972
6/24/1972
6/25/1972
6/26/1972
6/27/1972
Date
Passaic R. at Little Falls
Passaic R. at Millington
Passaic R. Near Chatham
Wanaque R. at Wanaque
Pequannock R. at Macopin Dam
Pompton R. at Pompton Plains
Rockaway above Boonton Reservoir
Rockaway below Boonton Reservoir
Whippany R. at Morristown

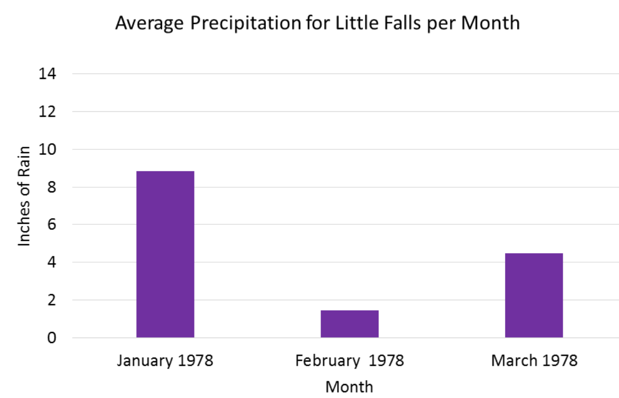

Average Precipitation for Northern New Jersey per Month
14
12
10
8
6
4
2
0
Inches
July 1975
August 1975
September 1975
Month


Average Precipitation for Little Falls per Month
14
12
10
8
6
4
2
0
Inches
July 1975
August 1975
September 1975
Month

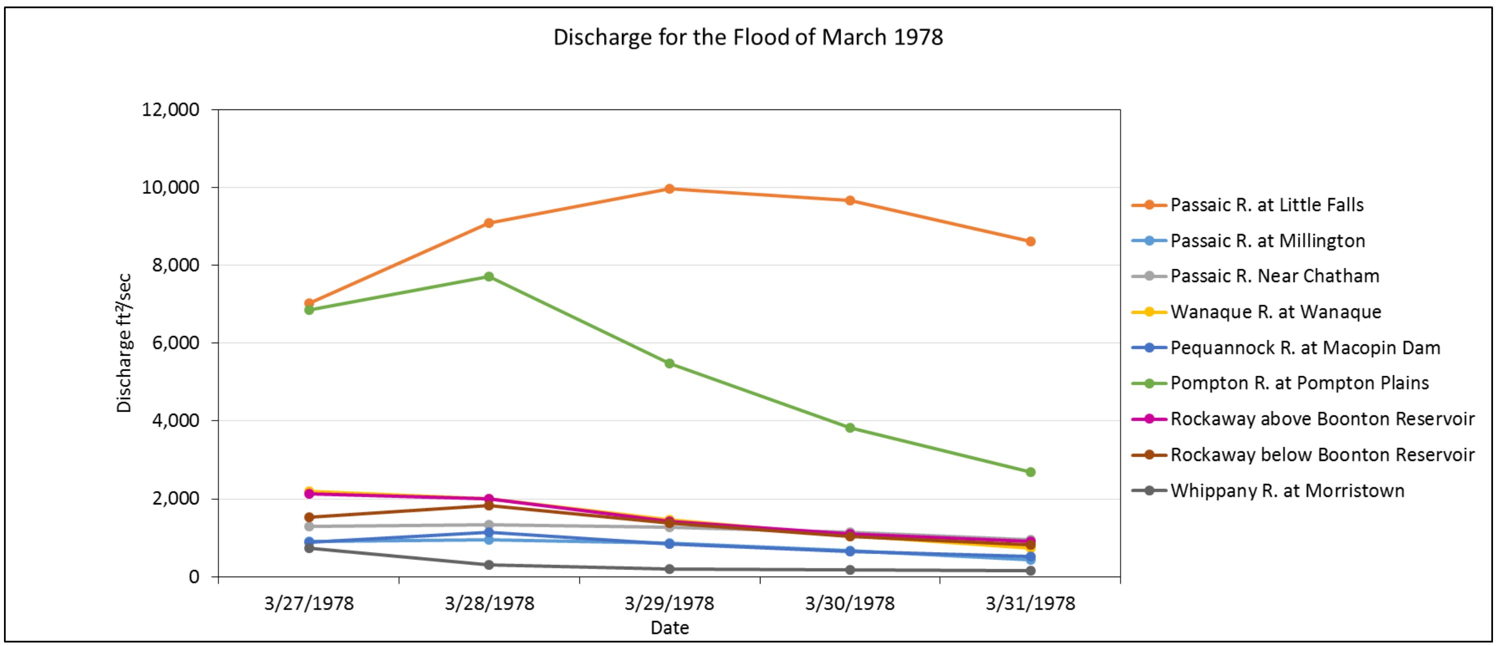

Discharge for the Flood of September 1975
12,000
10,000
8,000
6,000
4,000
2,000
0
Discharge ft³/sec
9/26/1975
9/27/1975
9/28/1975
9/29/1975
9/30/1975
Date
Passaic R. at Little Falls
Passaic R. at Millington
Passaic R. Near Chatham
Wanaque R. at Wanaque
Pequannock R. at Macopin Dam
Pompton R. at Pompton Plains
Rockaway above Boonton Reservoir
Rockaway below Boonton Reservoir
Whippany R. at Morristown



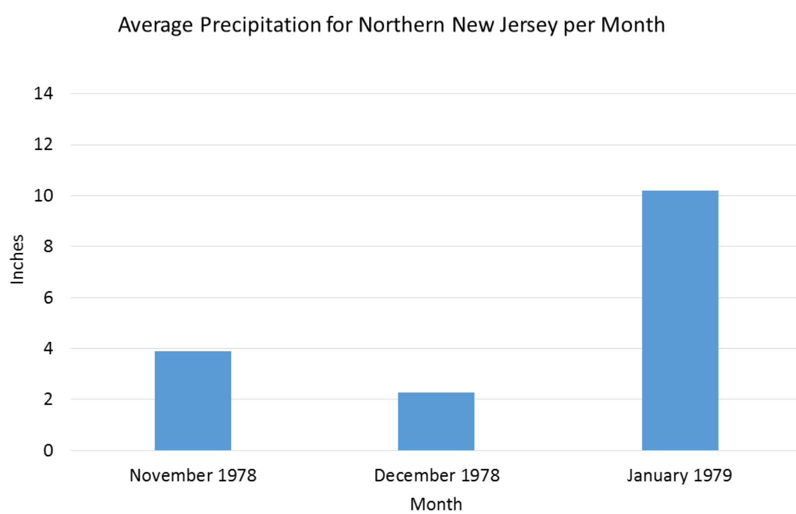
Average Precipitation for Northern New Jersey per Month
Inches
14
12
10
8
6
4
2
0
January 1978
February 1978
March 1978
Month

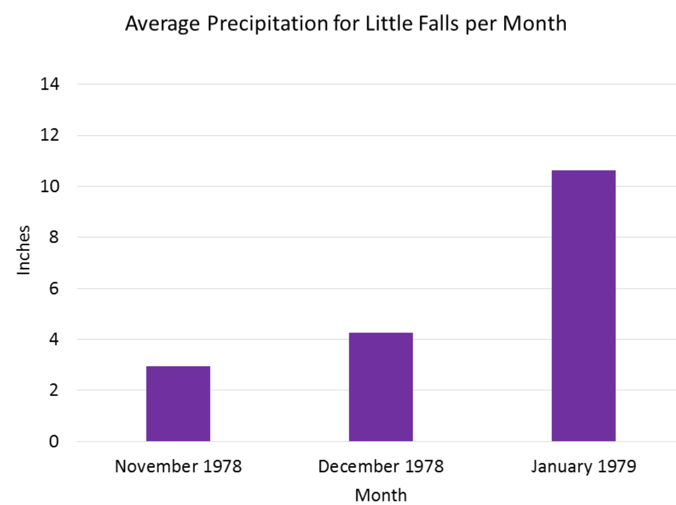


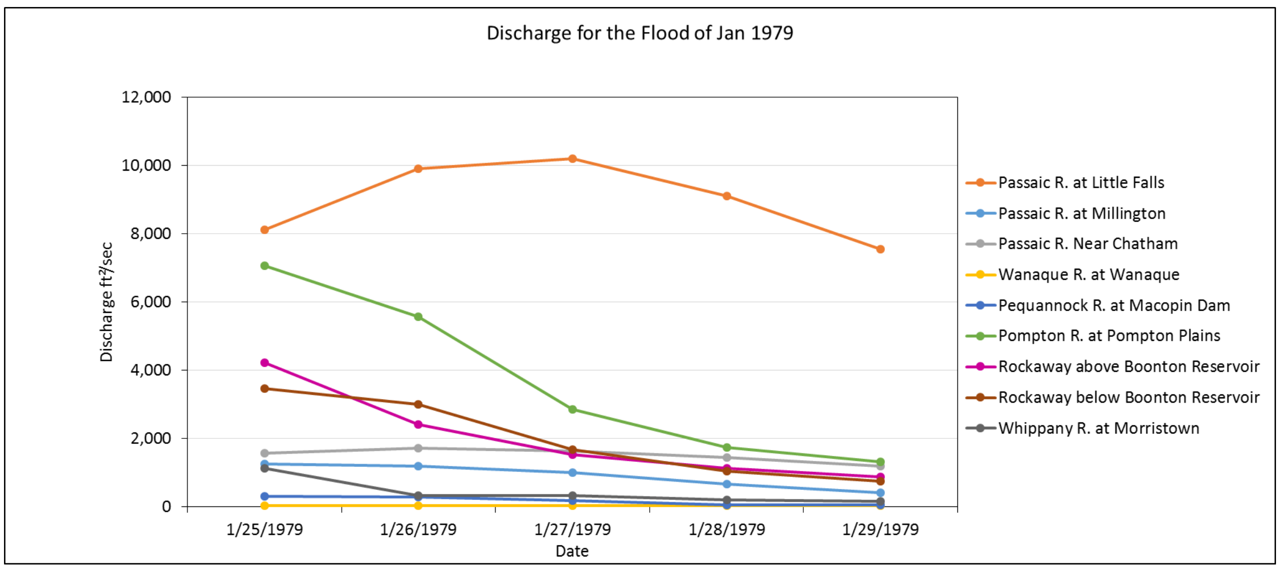
Average Precipitation for Little Falls per Month
Inches of Rain
14
12
10
8
6
4
2
0
January 1978
February 1978
March 1978
Month

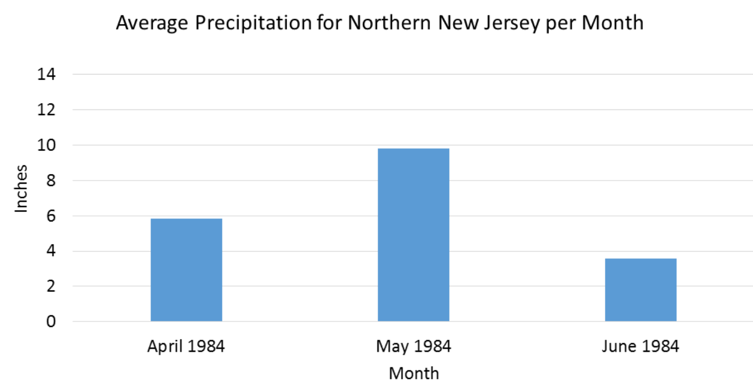


Discharge for the Flood of March 1978
Discharge ft³/sec
12,000
10,000
8,000
6,000
4,000
2,000
0
3/27/1978
3/28/1978
3/29/1978
3/30/1978
3/31/1978
Date
Passaic R. at Little Falls
Passaic R. at Millington
Passaic R. Near Chatham
Wanaque R. at Wanaque
Pequannock R. at Macopin Dam
Pompton R. at Pompton Plains
Rockaway above Boonton Reservoir
Rockaway below Boonton Reservoir
Whippany R. at Morristown

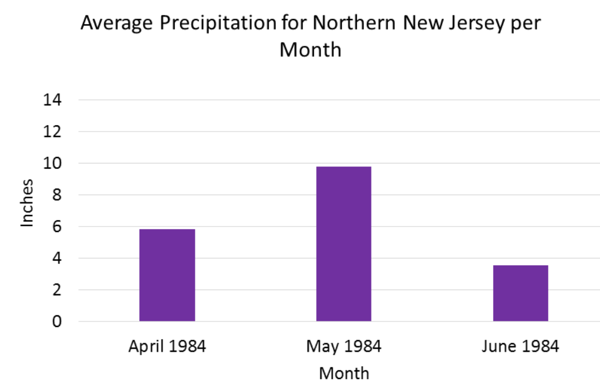

Average Precipitation for Northern New Jersey per Month
Inches
14
12
10
8
6
4
2
0
November 1978
December 1978
January 1979
Month


Average Precipitation for Little Falls per Month
Inches
14
12
10
8
6
4
2
0
November 1978
December 1978
January 1979
Month

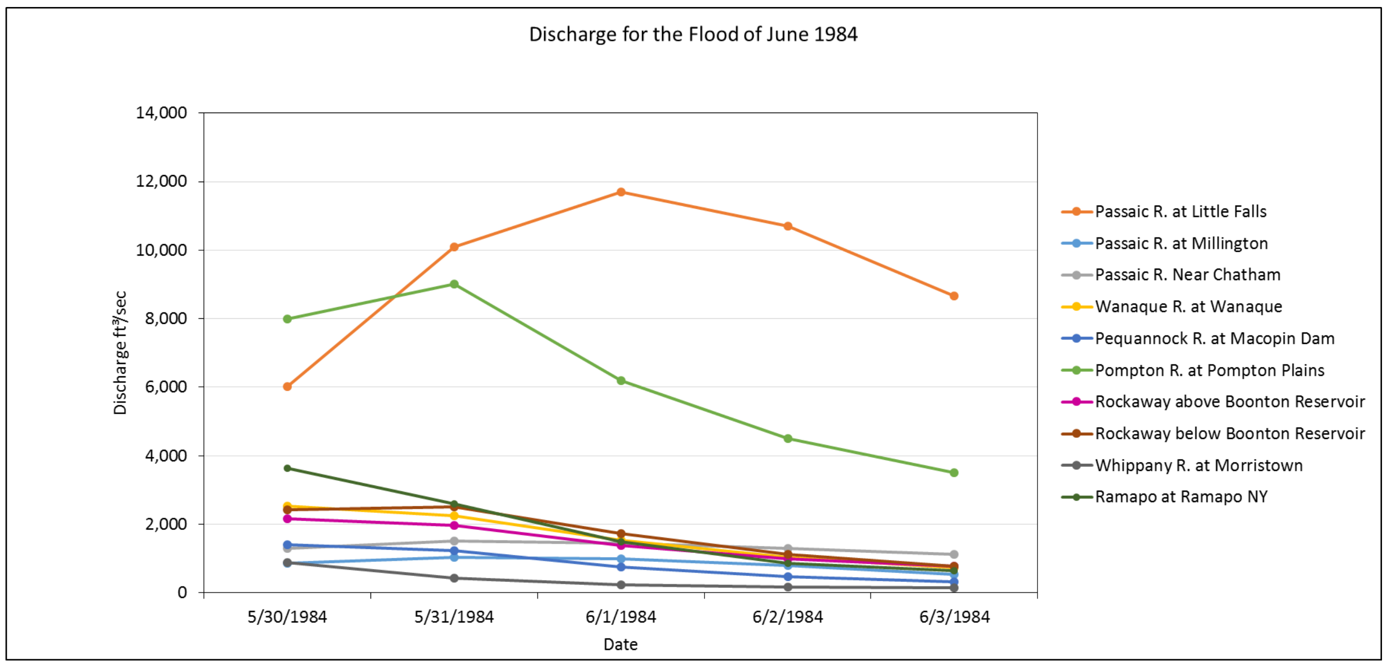

Discharge for the Flood of Jan 1979
Discharge ft²/sec
12,000
10,000
8,000
6,000
4,000
2,000
0
1/25/1979
1/26/1979
1/27/1979
1/28/1979
1/29/1979
Date
Passaic R. at Little Falls
Passaic R. at Millington
Passaic R. Near Chatham
Wanaque R. at Wanaque
Pequannock R. at Macopin Dam
Pompton R. at Pompton Plains
Rockaway above Boonton Reservoir
Rockaway below Boonton Reservoir
Whippany R. at Morristown



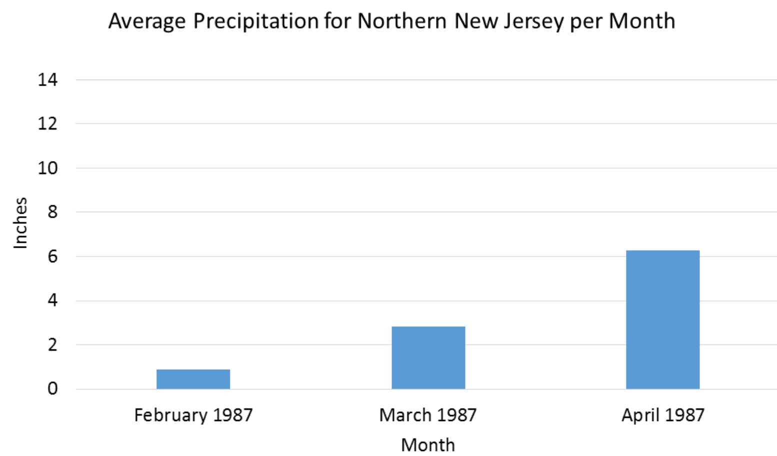
Average Precipitation for Northern New Jersey per Month
Inches
14
12
10
8
6
4
2
0
April 1984
May 1984
June 1984
Month

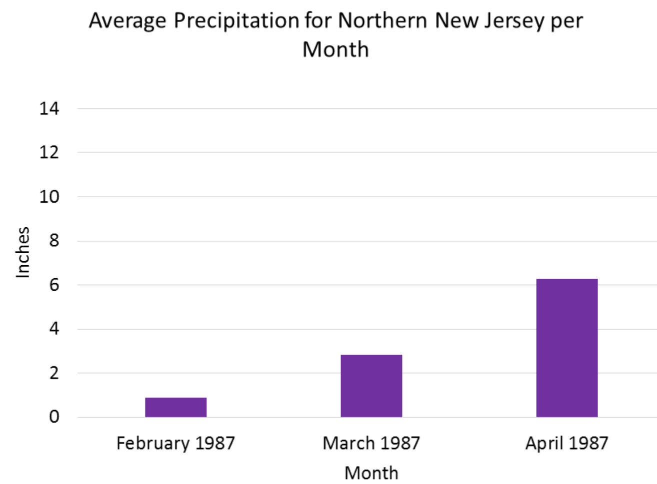


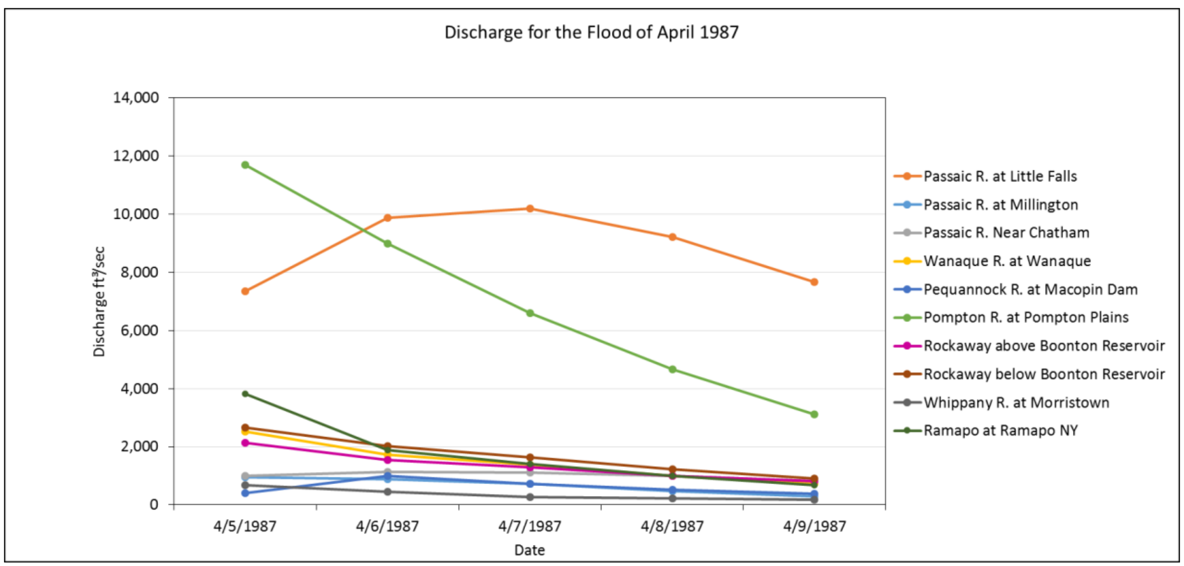
Average Precipitation for Northern New Jersey per Month
Inches
14
12
10
8
6
4
2
0
April 1984
May 1984
June 1984
Month

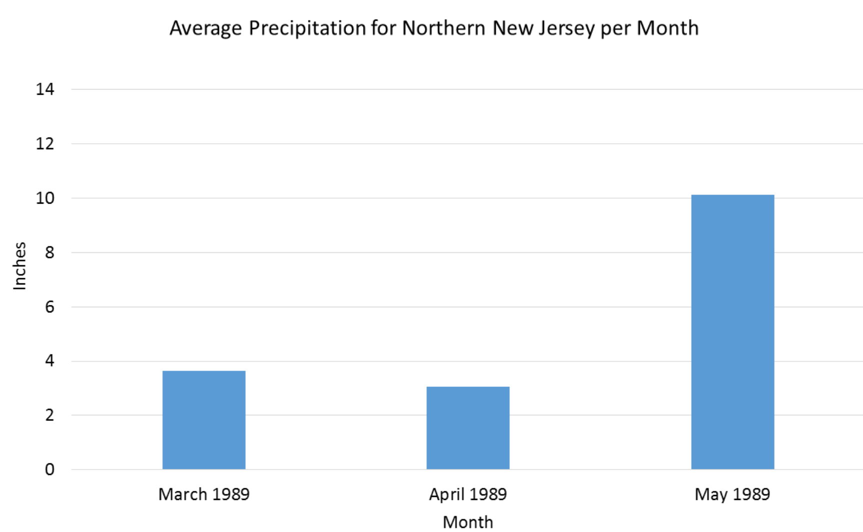


Discharge for the Flood of June 1984
Discharge ft³/sec
14,000
12,000
10,000
8,000
6,000
4,000
2,000
0
5/30/1984
5/31/1984
6/1/1984
6/2/1984
6/3/1984
Date
Passaic R. at Little Falls
Passaic R. at Millington
Passaic R. Near Chatham
Wanaque R. at Wanaque
Pequannock R. at Macopin Dam
Pompton R. at Pompton Plains
Rockaway above Boonton Reservoir
Rockaway below Boonton Reservoir
Whippany R. at Morristown
Ramapo at Ramapo NY

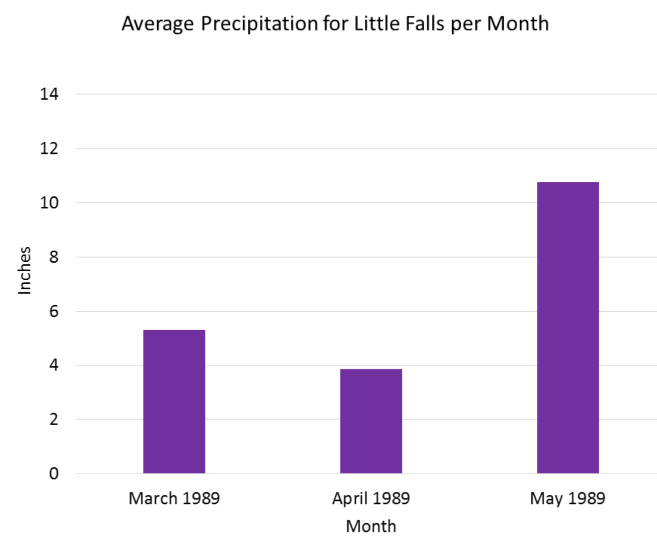

Average Precipitation for Northern New Jersey per Month
Inches
14
12
10
8
6
4
2
0
February 1987
March 1987
April 1987
Month


Average Precipitation for Northern New Jersey per Month
Inches
14
12
10
8
6
4
2
0
February 1987
March 1987
April 1987
Month

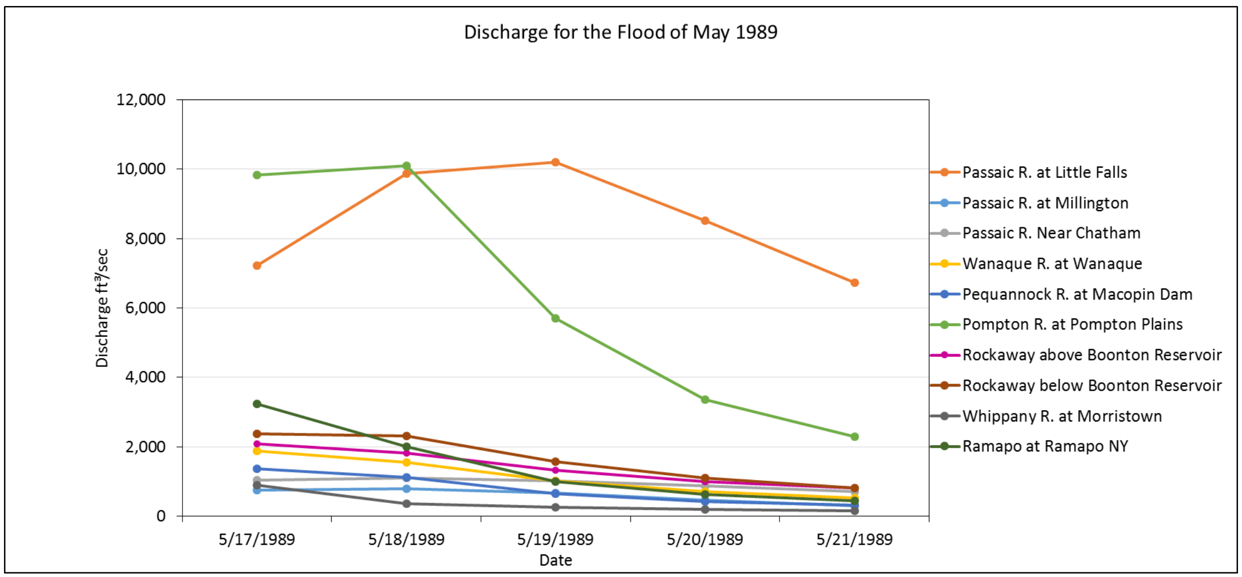

Discharge for the Flood of April 1987
Discharge ft³/sec
14,000
12,000
10,000
8,000
6,000
4,000
2,000
0
4/5/1987
4/6/1987
4/7/1987
4/8/1987
4/9/1987
Date
Passaic R. at Little Falls
Passaic R. at Millington
Passaic R. Near Chatham
Wanaque R. at Wanaque
Pequannock R. at Macopin Dam
Pompton R. at Pompton Plains
Rockaway above Boonton Reservoir
Rockaway below Boonton Reservoir
Whippany R. at Morristown
Ramapo at Ramapo NY



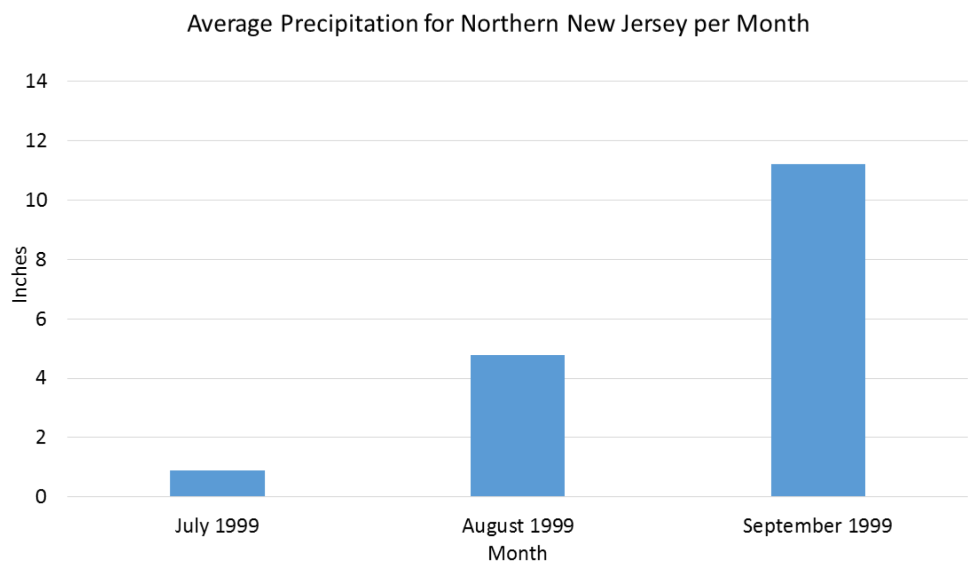
Average Precipitation for Northern New Jersey per Month
Inches
14
12
10
8
6
4
2
0
March 1989
April 1989
May 1989
Month



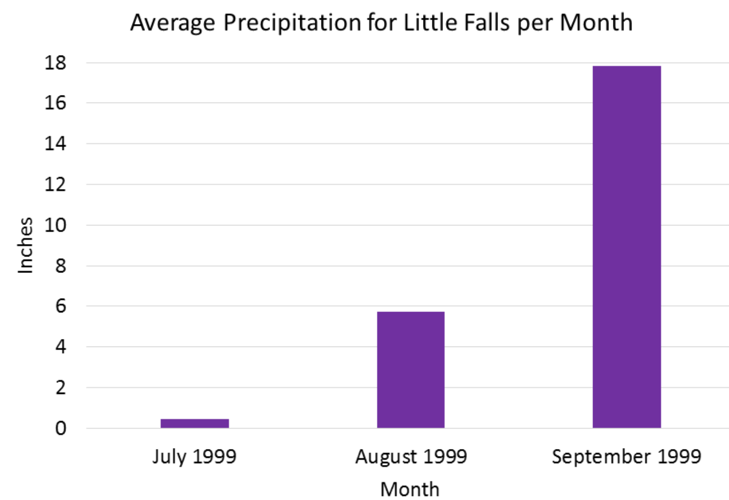
Average Precipitation for Little Falls per Month
Inches
14
12
10
8
6
4
2
0
March 1989
April 1989
May 1989
Month



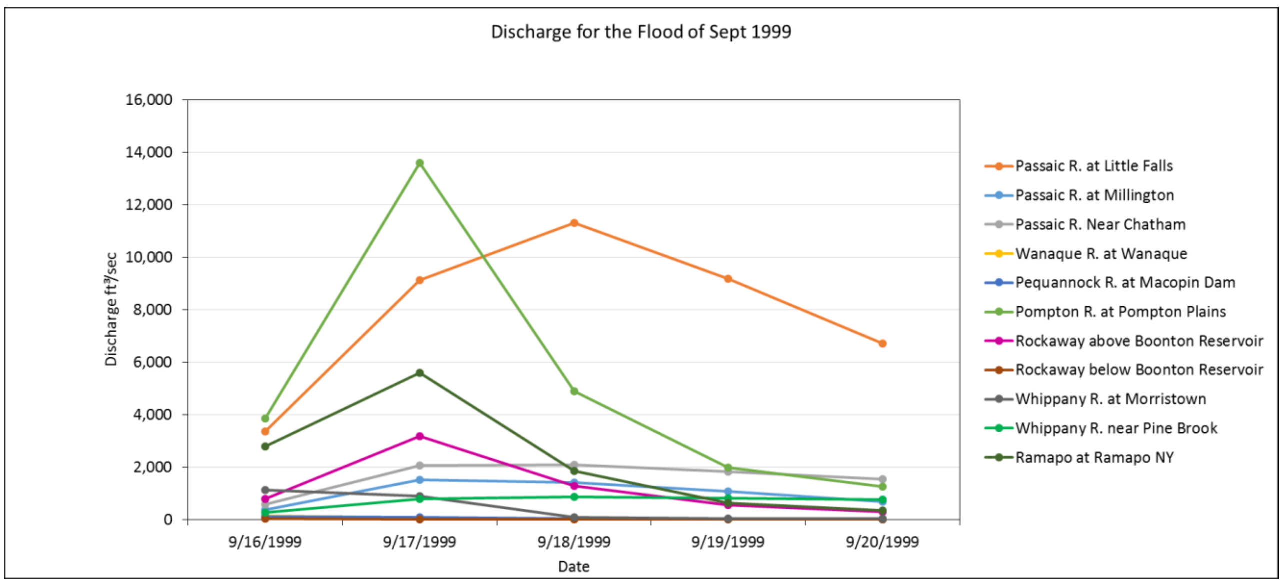
Discharge for the Flood of May 1989
Discharge ft³/sec
12,000
10,000
8,000
6,000
4,000
2,000
0
5/17/1989
5/18/1989
5/19/1989
5/20/1989
5/21/1989
Date
Passaic R. at Little Falls
Passaic R. at Millington
Passaic R. Near Chatham
Wanaque R. at Wanaque
Pequannock R. at Macopin Dam
Pompton R. at Pompton Plains
Rockaway above Boonton Reservoir
Rockaway below Boonton Reservoir
Whippany R. at Morristown
Ramapo at Ramapo NY



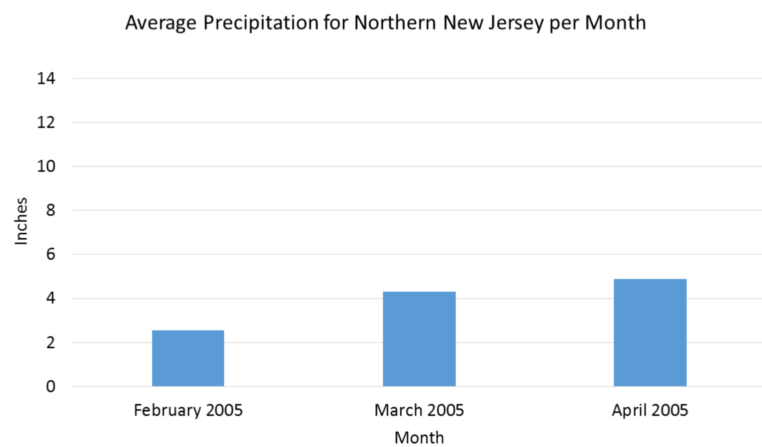
Average Precipitation for Northern New Jersey per Month
Inches
14
12
10
8
6
4
2
0
July 1999
August 1999
September 1999
Month

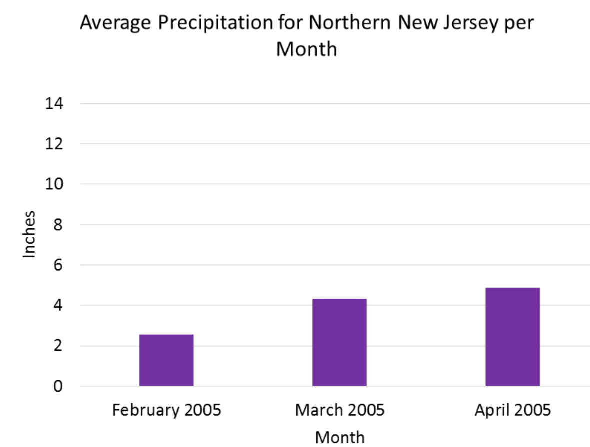


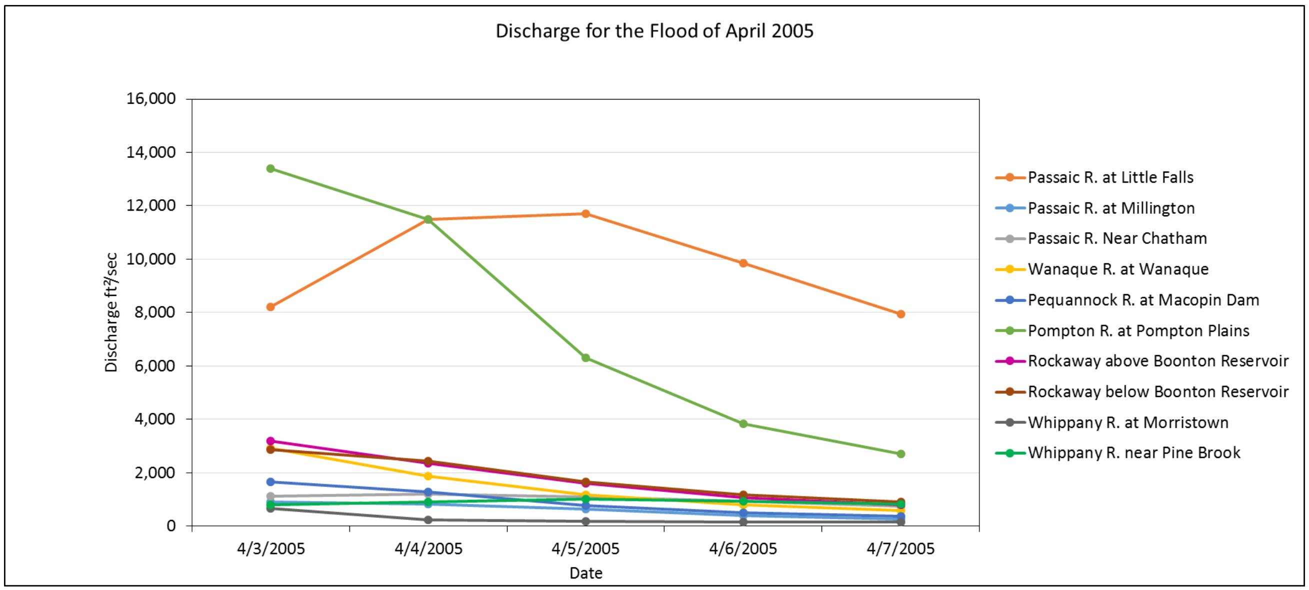
Average Precipitation for Little Falls per Month
Inches
18
16
14
12
10
8
6
4
2
0
July 1999
August 1999
September 1999
Month

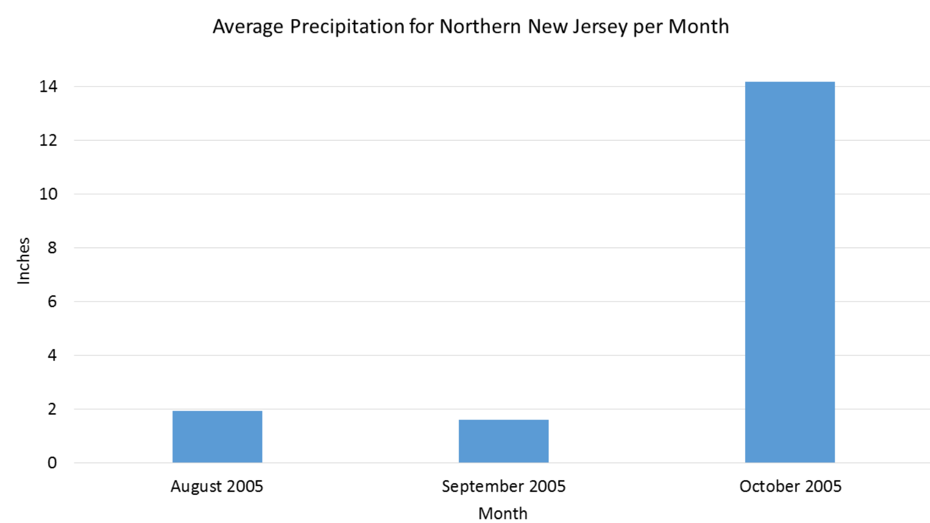


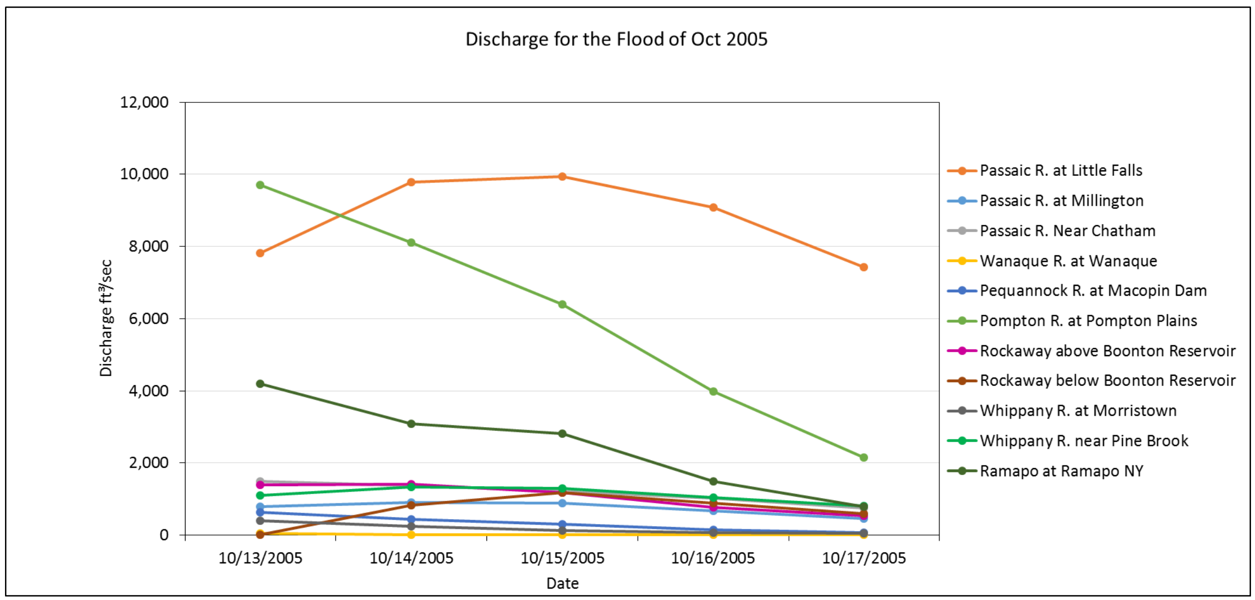
Discharge for the Flood of Sept 1999
Discharge ft³/sec
16,000
14,000
12,000
10,000
8,000
6,000
4,000
2,000
0
9/16/1999
9/17/1999
9/18/1999
9/19/1999
9/20/1999
Date
Passaic R. at Little Falls
Passaic R. at Millington
Passaic R. Near Chatham
Wanaque R. at Wanaque
Pequannock R. at Macopin Dam
Pompton R. at Pompton Plains
Rockaway above Boonton Reservoir
Rockaway below Boonton Reservoir
Whippany R. at Morristown
Whippany R. near Pine Brook
Ramapo at Ramapo NY

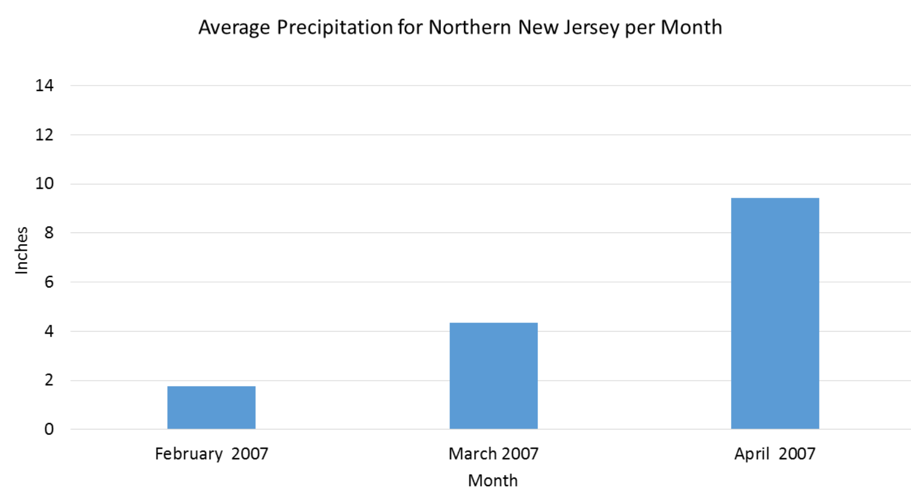

Average Precipitation for Northern New Jersey per Month
Inches
0
2
4
6
8
10
12
14
February 2005
March 2005
April 2005
Month

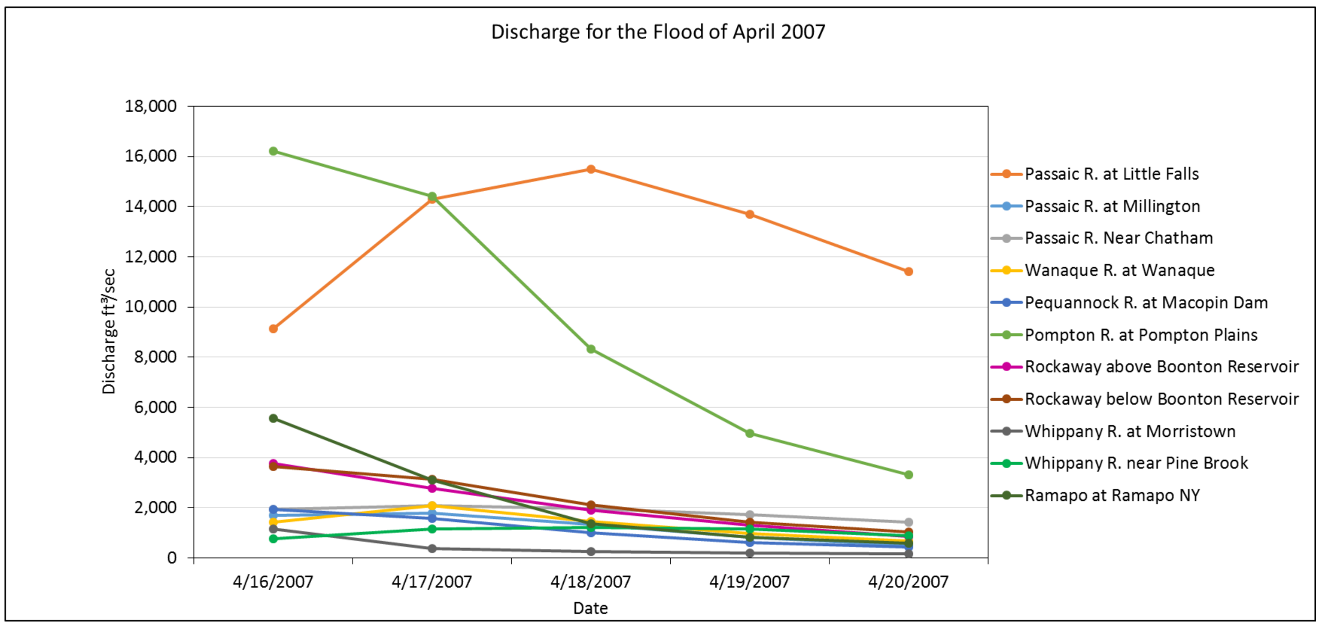

Average Precipitation for Northern New Jersey per Month
Inches
0
2
4
6
8
10
12
14
February 2005
March 2005
April 2005
Month

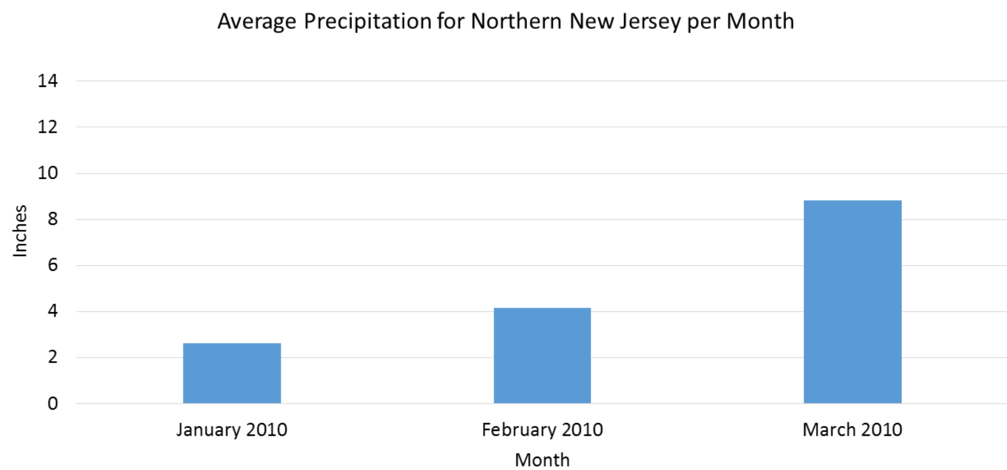

Discharge for the Flood of April 2005
Discharge ft³/sec
0
2,000
4,000
6,000
8,000
10,000
12,000
14,000
16,000
4/3/2005
4/4/2005
4/5/2005
4/6/2005
4/7/2005
Date
Passaic R. at Little Falls
Passaic R. at Millington
Passaic R. Near Chatham
Wanaque R. at Wanaque
Pequannock R. at Macopin Dam
Pompton R. at Pompton Plains
Rockaway above Boonton Reservoir
Rockaway below Boonton Reservoir
Whippany R. at Morristown
Whippany R. near Pine Brook

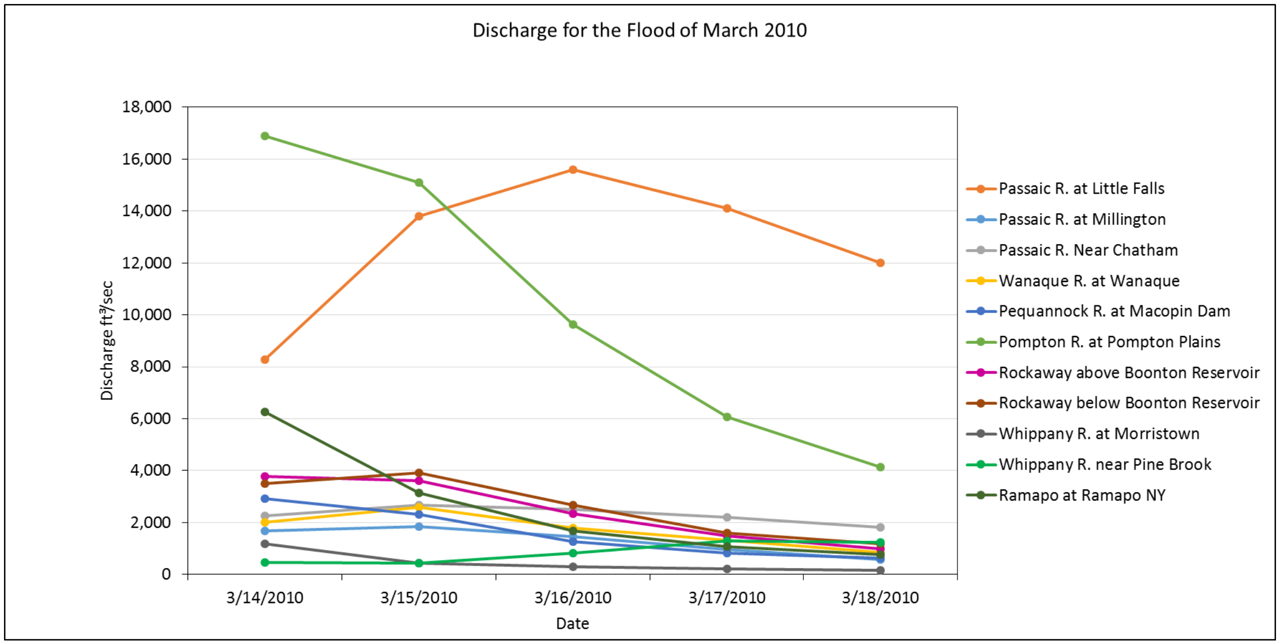

Average Precipitation for Northern New Jersey per Month
Inches
0
2
4
6
8
10
12
14
August 2005
September 2005
October 2005
Month

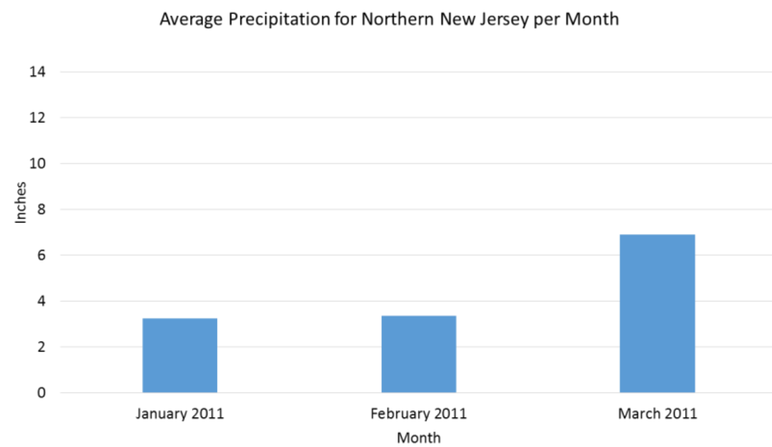

Discharge for the Flood of Oct 2005
Discharge ft³/sec
0
2,000
4,000
6,000
8,000
10,000
12,000
10/13/2005
10/14/2005
10/15/2005
10/16/2005
10/17/2005
Date
Passaic R. at Little Falls
Passaic R. at Millington
Passaic R. Near Chatham
Wanaque R. at Wanaque
Pequannock R. at Macopin Dam
Pompton R. at Pompton Plains
Rockaway above Boonton Reservoir
Rockaway below Boonton Reservoir
Whippany R. at Morristown
Whippany R. near Pine Brook
Ramapo at Ramapo NY

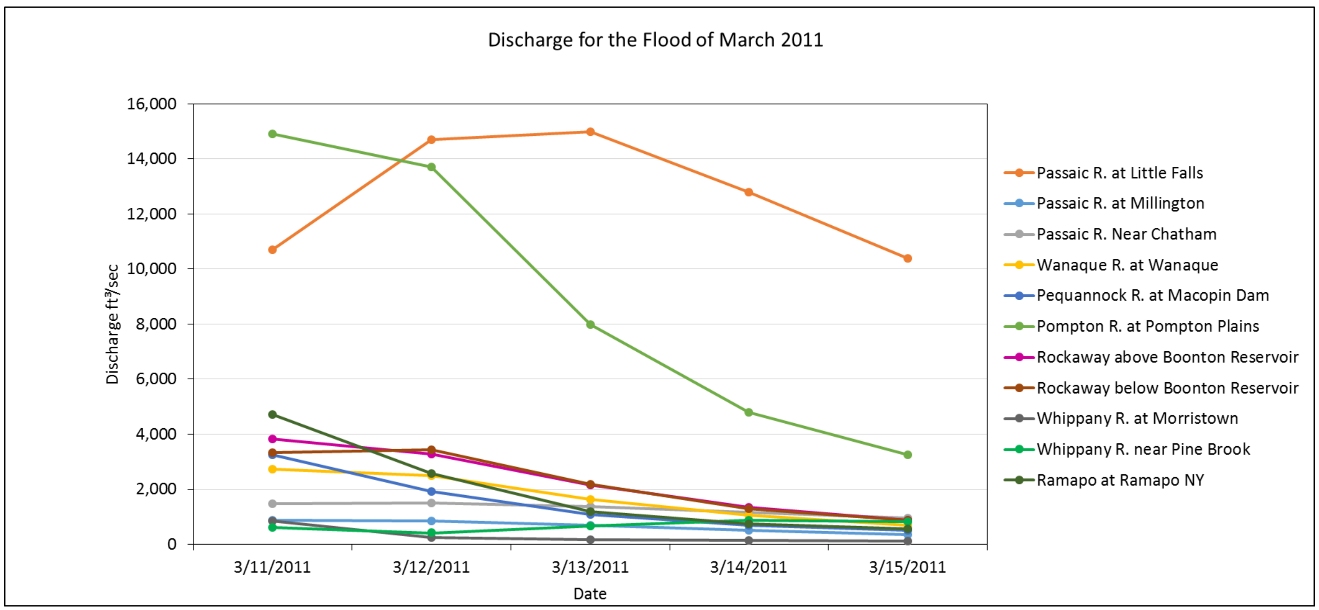

Average Precipitation for Northern New Jersey per Month
Inches
14
12
10
8
6
4
2
0
February 2007
March 2007
April 2007
Month

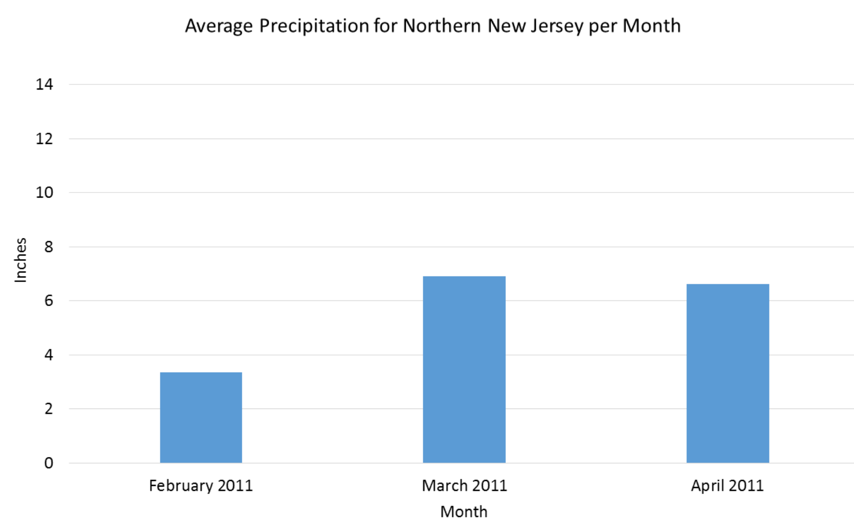

Discharge for the Flood of April 2007
Discharge ft³/sec
18,000
16,000
14,000
12,000
10,000
8,000
6,000
4,000
2,000
0
4/16/2007
4/17/2007
4/18/2007
4/19/2007
4/20/2007
Date
Passaic R. at Little Falls
Passaic R. at Millington
Passaic R. Near Chatham
Wanaque R. at Wanaque
Pequannock R. at Macopin Dam
Pompton R. at Pompton Plains
Rockaway above Boonton Reservoir
Rockaway below Boonton Reservoir
Whippany R. at Morristown
Whippany R. near Pine Brook
Ramapo at Ramapo NY

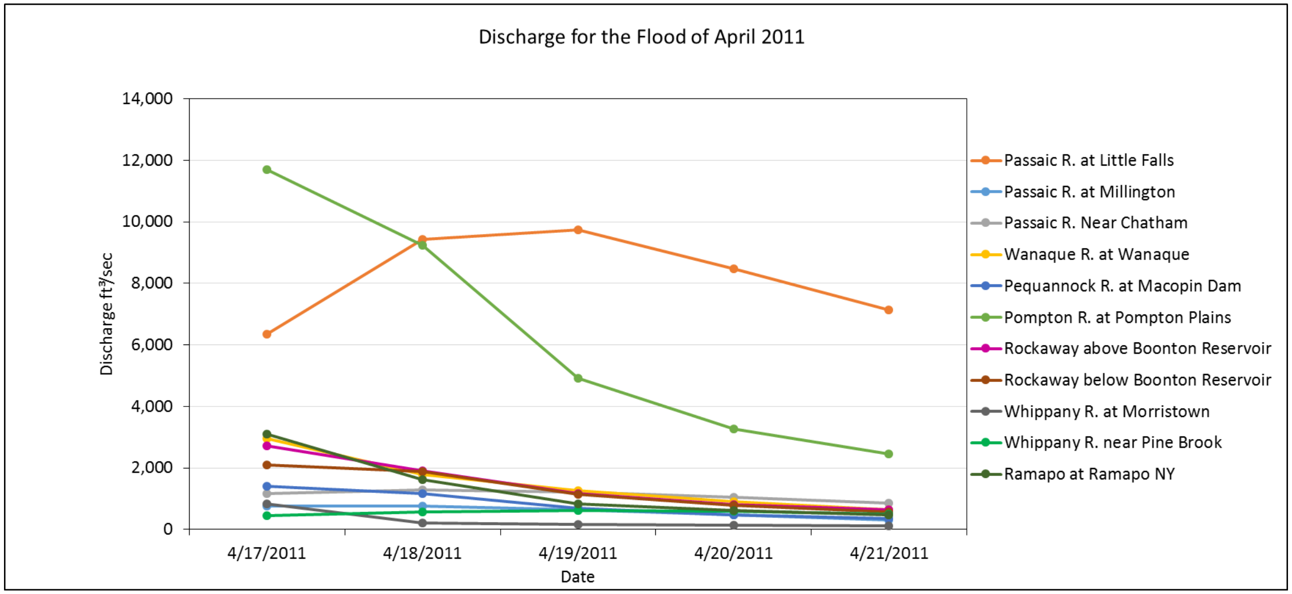

Average Precipitation for Northern New Jersey per Month
Inches
14
12
10
8
6
4
2
0
January 2010
February 2010
March 2010
Month

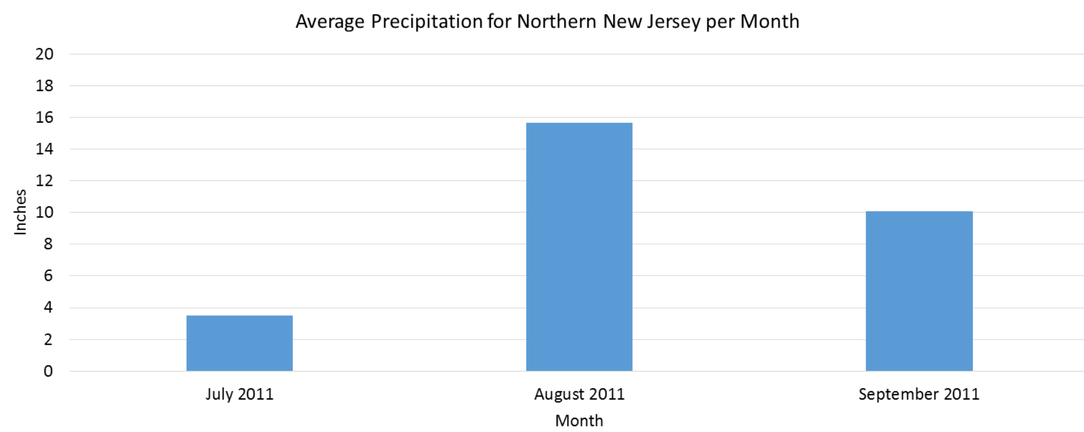

Discharge for the Flood of March 2010
Discharge ft³/sec
18,000
16,000
14,000
12,000
10,000
8,000
6,000
4,000
2,000
0
3/14/2010
3/15/2010
3/16/2010
3/17/2010
3/18/2010
Date
Passaic R. at Little Falls
Passaic R. at Millington
Passaic R. Near Chatham
Wanaque R. at Wanaque
Pequannock R. at Macopin Dam
Pompton R. at Pompton Plains
Rockaway above Boonton Reservoir
Rockaway below Boonton Reservoir
Whippany R. at Morristown
Whippany R. near Pine Brook
Ramapo at Ramapo NY

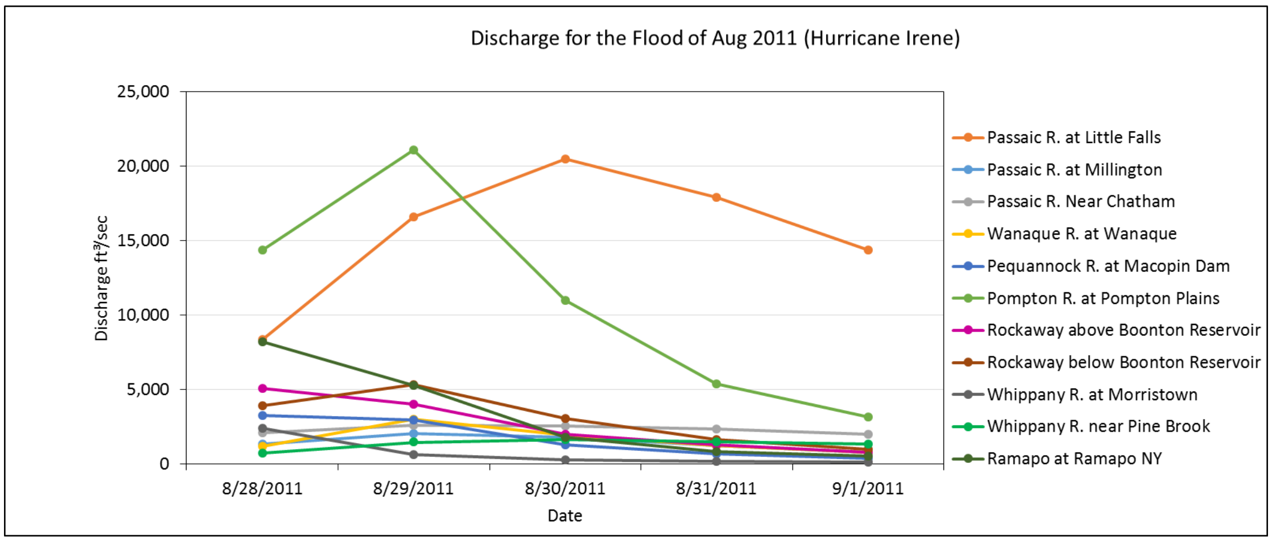

Average Precipitation for Northern New Jersey per Month
14
12
10
8
6
4
2
0
Inches
January 2011
February 2011
March 2011
Month

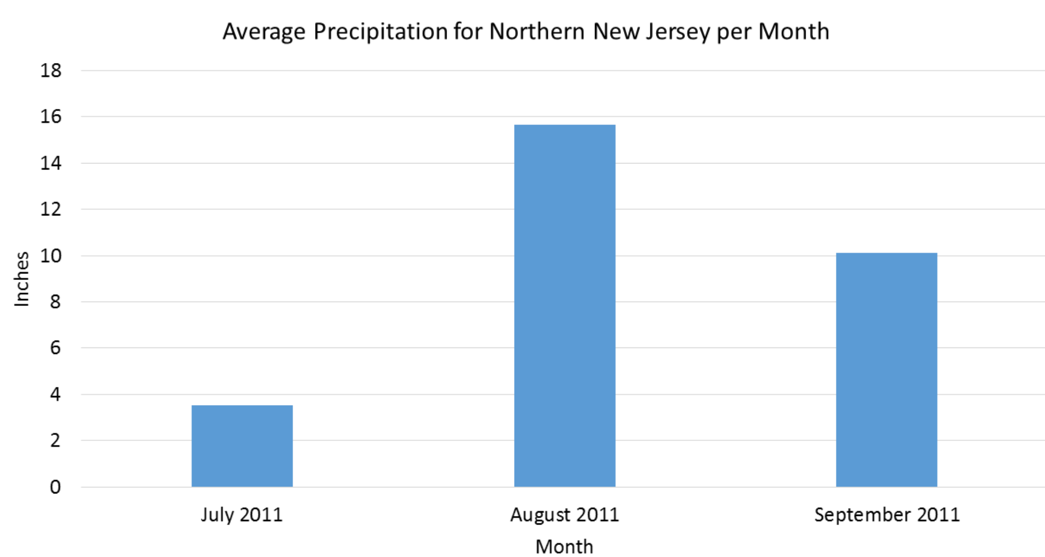

Discharge for the Flood of March 2011
16,000
14,000
12,000
10,000
8,000
6,000
4,000
2,000
0
Discharge ft³/sec
3/11/2011
3/12/2011
3/13/2011
3/14/2011
3/15/2011
Date
Passaic R. at Little Falls
Passaic R. at Millington
Passaic R. Near Chatham
Wanaque R. at Wanaque
Pequannock R. at Macopin Dam
Pompton R. at Pompton Plains
Rockaway above Boonton Reservoir
Rockaway below Boonton Reservoir
Whippany R. at Morristown
Whippany R. near Pine Brook
Ramapo at Ramapo NY


Average Precipitation for Northern New Jersey per Month
Inches
14
12
10
8
6
4
2
0
February 2011
March 2011
April 2011
Month

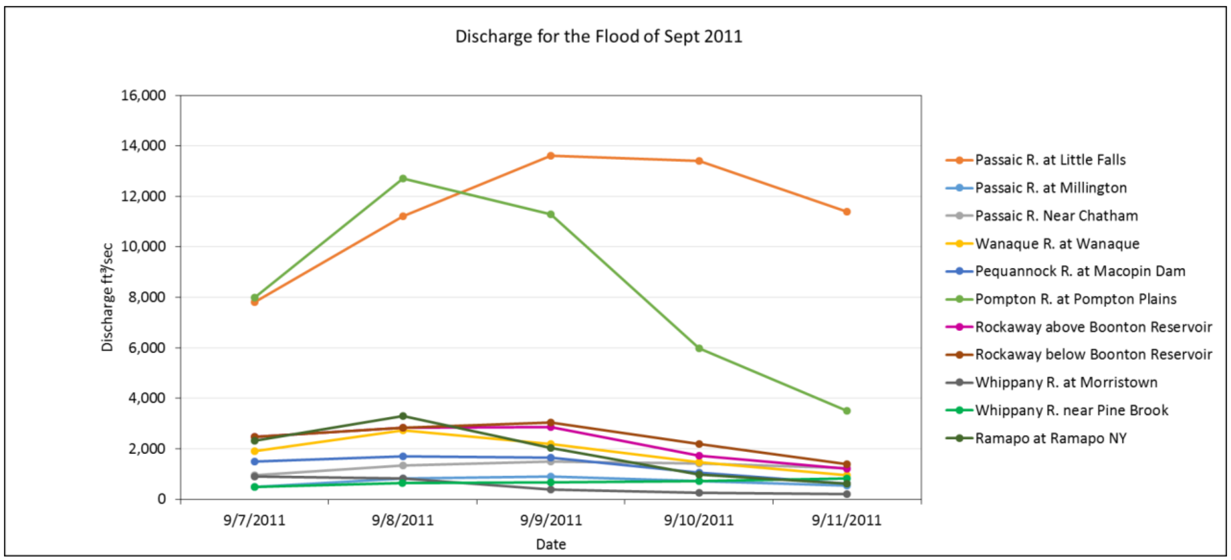

Discharge for the Flood of April 2011
Discharge ft³/sec
14,000
12,000
10,000
8,000
6,000
4,000
2,000
0
4/17/2011
4/18/2011
4/19/2011
4/20/2011
4/21/2011
Date
Passaic R. at Little Falls
Passaic R. at Millington
Passaic R. Near Chatham
Wanaque R. at Wanaque
Pequannock R. at Macopin Dam
Pompton R. at Pompton Plains
Rockaway above Boonton Reservoir
Rockaway below Boonton Reservoir
Whippany R. at Morristown
Whippany R. near Pine Brook
Ramapo at Ramapo NY


Average Precipitation for Northern New Jersey per Month
Inches
20
18
16
14
12
10
8
6
4
2
0
July 2011
August 2011
September 2011
Month


Discharge for the Flood of Aug 2011 (Hurricane Irene)
Discharge ft³/sec
25,000
20,000
15,000
10,000
5,000
0
8/28/2011
8/29/2011
8/30/2011
8/31/2011
9/1/2011
Date
Passaic R. at Little Falls
Passaic R. at Millington
Passaic R. Near Chatham
Wanaque R. at Wanaque
Pequannock R. at Macopin Dam
Pompton R. at Pompton Plains
Rockaway above Boonton Reservoir
Rockaway below Boonton Reservoir
Whippany R. at Morristown
Whippany R. near Pine Brook
Ramapo at Ramapo NY



Average Precipitation for Northern New Jersey per Month
Inches
18
16
14
12
10
8
6
4
2
0
July 2011
August 2011
September 2011
Month



Discharge for the Flood of Sept 2011
Discharge ft³/sec
16,000
14,000
12,000
10,000
8,000
6,000
4,000
2,000
0
9/7/2011
9/8/2011
9/9/2011
9/10/2011
9/11/2011
Date
Passaic R. at Little Falls
Passaic R. at Millington
Passaic R. Near Chatham
Wanaque R. at Wanaque
Pequannock R. at Macopin Dam
Pompton R. at Pompton Plains
Rockaway above Boonton Reservoir
Rockaway below Boonton Reservoir
Whippany R. at Morristown
Whippany R. near Pine Brook
Ramapo at Ramapo NY