Sea Level, Tidal, and River Flow Trends in the Lower Columbia River Estuary, 1853-Present

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Sea level, tidal and river flow trends in the Lower Columbia River Estuary, 1853-present

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Key Points

- Extensive data archaeology and digitization of analog records has been used to produce high resolution data from 1853-1876 at Astoria
- Relative sea-level rise since the 1850s is 0.06m; we estimate a geocentric rise of 0.11m, after accounting for vertical land motion.
- Significant variations in seasonal sea level and tides are observed due to altered river flow; a secular increase of 0.1 m in tidal range is observed.
Abstract

Few tidal records are available pre-1900 for the Pacific Ocean. We improve data coverage by recovering historical tabulations and digitizing analog tide rolls from Astoria, Oregon for 1853-1876. Nearly 13,500 overlapping images of tides from 1855-1870 were digitized at a 6 minute resolution using a line-finding algorithm. Available hourly and high/low tabulations were also digitized, as were nearby hourly records from 1933-1943. Uncertainty was assessed by evaluating manual staff measurements, historical documents, and leveling surveys. Results suggest that uncertainty in mean sea level varies from ±0.07m (early 1850s) to ±0.03m (1867-1876) and is driven primarily by datum and benchmark uncertainty, rather than measurement precision, data reduction procedures, or hydrodynamic changes. We also corrected an up-to-0.05m error in the 1925-1960 tidal datum at Astoria. Harmonic analysis shows that major tidal constituents increased by up to 7% between 1855 and 2018. Mean tidal range increased by 0.1m (5%), with more change occurring in July (0.17m larger than winter (0.07m larger). By contrast, sea level increased most in winter, and least in spring/summer. Tidally-based estimates of river discharge suggest that these observations are caused by a ~50% reduction in peak spring discharge and a 30-60% increase in winter discharge. No evidence of altered upwelling is found. Overall, Astoria relative sea level (RSL) increased by 0.06m ±0.04m since the 1858-1876 epoch, or, after accounting for vertical land motion, 0.11 ± 0.09m. Consistent with GNSS measurements, RSL has dropped near the estuary mouth since 1905, indicating a strong tectonic influence.

1.0 Introduction

Tidal records form the earliest instrumental data we have for assessing long-term hydrodynamic and sea level changes in coastal regions, and archival records have been used to assess trends in storminess (Bromirski et al., 2003), river flow (Moftakhari et al., 2013), storm surge (Woodworth and Blackman, 2002; Talke et al., 2014), tides (Ray, 2006; Jay 2009; Woodworth, 2010; Haigh et al., 2020; Talke & Jay, 2020), seasonal sea level variability (Dangendorf et al., 2013), and sea-level rise and acceleration (e.g., Jevrejeva et al. 2008; Church & White, 2011; Hogarth 2014; Hay et al. 2015). However, one of the challenges in interpreting historical variability and trends in both high-frequency (e.g., hourly) and mean water levels (e.g., annual) is the paucity of gauge data before 1900 and its bias towards Northern Europe (e.g., Holgate et al., 2013, Woodworth et al., 2017). Since many thousands of additional station years exist in undigitized form (Pouvreau 2008; Bradshaw et al., 2015; Talke & Jay, 2013, 2017), data rescue
efforts can potentially address this need. Hence, many recent efforts have focused on finding, digitizing and reanalyzing historical tidal records, and tying them to a stable datum (e.g., Burgette et al. (2009), Testut et al. (2010), Bradshaw et al. (2016), Marcos et al. (2011,2013), Wöppelmann et al., (2008,2014) and Talke et al., (2014, 2018)). Here, we digitize a long 19th century record from Astoria, Oregon, from 1853-1876, a period for which only one other high frequency, hourly measurement (from San Francisco) is currently available for the entire Pacific Ocean. By digitizing at hourly frequency, we are able to characterize many factors that influence water level in an estuary, including tides and river discharge, thereby improving interpretation of variability and trends in sea level.

In the Eastern Pacific, coupled oceanic and hydro-meteorological forcing produces a variable response in relative sea level (RSL) over multiple time scales. Summertime upwelling winds and wintertime downwelling winds cause spatial, seasonal and interannual variability in west coast RSL (Chelton & Davis, 1982; Strub et al., 1987). Similarly, El Niño conditions produce interannual variability in sea level and a spatially variable sea level anomaly that decreases from the equator northwards (e.g., Hamlington et al., 2015). Shifts in Pacific Basin trade winds depressed sea-level rise rates in the eastern Pacific between roughly 1980-1990 and 2010 (e.g., Merrifield 2011, Bromirski et al. 2011), but may have recently reversed and produced elevated rise rates (e.g., Hamlington et al., 2016; Merrifield &Thomson, 2018). Additionally, vertical land motion caused by glacial isostatic adjustment and tectonics cause significant (order 1mm/year) variations in RSL rise, often over length scales as small as 10km (Burgette et al., 2009; NRC, 2012). Similarly, groundwater extraction and other local factors can lead to large local variability; for example, RSL rise varies by up to 10 mm/yr in the San Francisco Bay Area (Shirzaei & Bürgmann, 2018). Channel deepening can reduce the barotropic slope within rivers and estuaries, feeding back into mean-water level measurements (Jay et al., 2011; Ralston et al., 2019). Finally, long-term annual and seasonal trends in river flow (e.g., Naik & Jay, 2011; Moftakhari et al., 2013) can affect the local sea level in an estuary and leave an imprint on coastal RSL variability (Piecuch et al. 2018).
In this paper, we use archival research and data recovery to evaluate the spatial and temporal variability of water level in the Lower Columbia River Estuary (see map in Figure 1). The mid-19th century records from Astoria predate most regional industrial and agricultural development and provide a glimpse of how water levels (tides and sea level) in the Columbia River Estuary (CRE) varied before navigational improvements (e.g., channel deepening) and river regulation/flood control. Because many paper-based tabulations of water level were missing or only available at the times of high and low water, we extracted hourly records from the original tide rolls (also called marigrams or mareograms; see Talke & Jay, 2013, and Figure 2). Many additional local water level and meta-data records were recovered, digitized, and assessed to corroborate results and aid interpretation. Mean-water level from more seaward locations such as Fort Stevens (1905-2014, intermittent) and Youngs Bay (1931-1943) were also recovered to provide insights into spatial variability.

We use the reconstructed data set to evaluate the biases, uncertainty and variability introduced by altered gauge location, vertical land motion, datum errors, and trends in river flow. To help estimate the hydrodynamic gradient between the modern tide gauge at Tongue Point (1925-present) and the historical gauge in Astoria, we installed and have maintained a radar tide gauge at the historical gauge location since October 2015, and tied it to local, long term benchmarks. We use daily staff/gauge comparisons from 1925-1960 to update the Burgette et al. (2009) conclusion that the Tongue Point station datum was unstable pre-1950, leading to an underestimation of sea-level rise by ~0.05m. The combination of hydrodynamic and staff datum corrections improve the comparison between coincident Tongue Point and Youngs Bay data, justifying the approach. Finally, we demonstrate large, tectonically driven variability in local mean water levels. At Fort Stevens near the coast, RSL decreased between 1905 to 2014, while around Astoria the local sea level rose by ~0.06 ± 0.04m since the mid-1800s (see Figure 1 for locations). This highly variable hydrodynamic and tectonic environment suggests that such regions need to be monitored with a denser array of gauges that is commensurate with local variability, to help accurately assess the causes and effects of future RSL rise.
2.0 Methods

In the following sections we provide a brief history of tidal measurements in the lower Columbia River (section 2.1), describe our digitization method (section 2.2), detail historical leveling and our datum reconstruction of 19th and early 20th century measurements (section 2.3), and describe how we estimate river discharge from 1855 to 1876 using available tidal, river stage, and river discharge records. These steps enable an understanding of how tides, river discharge, and sea-level have changed from the 1853-1876 period to the present (see Results). For reference, Table 1 provides a synopsis of all the archival water level records we found, used, and/or digitized. Additional details and examples of archival records are provided in the electronic supplement, and a link to data records are provided in the acknowledgement section.

2.1 Historical Setting and Measurements

The Columbia River, the fourth largest river in North America, currently discharges at an annual average rate of ~7,100 m$^3$/s to the Pacific Ocean (1970-1999 period; Naik & Jay, 2011). The mesotidal estuary exhibits mixed, semi-diurnal tides with a mean tidal range at Astoria Tongue Point (National Oceanographic and Atmospheric Administration (NOAA) gauge 9439040) of 2.06m (1983-2001 epoch) and a great diurnal range of 2.62m. At the Columbia River mouth, installation of a system of jetties beginning in the 1880s narrowed the entrance from 9 to 3.5km and deepened the shipping channel from 8 to 15-17m. The river has also been lengthened; Point Adams, near the historical mouth, is now at River Kilometer 10 (Rkm 10; We use the coordinate system of the US Army Corps of Engineers, which has its origin at the modern river mouth and is positive in the upstream direction). Since the late 1800s, the channel between the coast and the Portland metropolitan area (Rkm 170; Figure 1) was deepened from 5-8m to 13m, intertidal areas were filled, dredge spoil islands were created, the channel length reduced, and the hydraulic efficiency increased (Sherwood et al., 1990; Helaire et al., 2019). Land reclamation removed habitat, while pile dikes narrowed the channel width and altered channel/shoal dynamics (Sherwood et al. 1990). Large-scale water diversion began in the 1890s (Naik & Jay, 2005) and now accounts for about 7-8% of the total flow (Naik & Jay, 2011). Finally, hydropower generation, flood control, deforestation, and long-term changes in climate have altered the timing and magnitude of river flow: the peaks of annual spring time floods (freshets)
have been reduced by 45% and occur up to a month earlier than in the late 19\textsuperscript{th} century, while late summer and winter flows have increased due to hydropower production (Matheussen et al., 2000; Naik & Jay, 2005, 2011).

The recent re-discovery of extensive water level measurements in the Lower Columbia River Estuary (LCRE) (Table 1; Talke & Jay, 2013; Helaire et al., 2019; Talke & Jay, 2017) provides the opportunity to assess empirically how water levels, tides, and river flow have changed over the past 165 years, since the mid-19\textsuperscript{th} century. Multiple types of water level data from between 1853 to 1972 were recovered from five different archives by taking photographs of documents, and manually digitized (Table 1). Meta-data from seven archives was also consulted to interpret the measurements, and included letters, observer notes, leveling surveys, and summary sheets. Details regarding data types and the archives from which they were obtained are included in Table 1, and examples of archival data and meta data are shown in the Supplement.

2.1.1 Pre-1925 measurements

Systematic tidal measurements within the Lower Columbia River Estuary (LCRE) began soon after the 1846 Oregon Treaty formalized US ownership of the Oregon and Washington territories (see Table 1; earlier measurements by the US or British Navy possibly occurred, but have not been found). The US Coast Survey organized several short tidal surveys of the LCRE in 1850 and 1852, and began continuous measurements with an automatic gauge in the town of Astoria in July 1853. The gauge was run continuously until October 31\textsuperscript{st}, 1876, then moved to Sausalito, California. The US Army Corps of Engineers made tidal measurements in 1883,1884, and from 1886 or 1887 until 1899 (Talke & Jay, 2013). However, only the monthly averages from 1883 and 1884 have been found. Army Corps records indicate that the automatic gauge was moved in 1899 to Fort Stevens, Oregon (Rkm 11; see Figure 1), and measurements were continued until at least 1907. The measurements from Fort Stevens comprised one of 26 mean sea level records used to define the National Geodetic Vertical Datum of 1929 (NGVD-29 datum; see Schomaker & Berry, 1981), the predecessor to the currently used North American Vertical Datum of 1988
(NAVD-88). However, despite its historical significance, no high-resolution records from Fort Stevens have yet been found, and only the mean sea level for 1905-1906 was recovered, from a summary sheet obtained from NOAA (see Supplement Figure S.2.28). More details documenting the history of tidal measurements can be found in Supplement Section S.1.

2.1.2 Post-1925 measurements

A period of 49 years and 3 months passed between the end of the Coastal Survey data set in 1876 and the begin of the modern Coast and Geodetic Survey (now NOAA) gauge at Astoria Tongue Point (1925-present; station 9439040). Daily discharge measurements began upstream of the head of tides in 1878 (USGS station 14105700), and once-a-day stage measurements began at Portland, Oregon by January 1876 (Table 1; Wilson, 1878). Additional tabulations of water level at Vancouver, Washington, from 1872-1877 (spring freshet only) were found in notebooks of the US Coast Survey from their 1877 hydrographic survey of the LCRE. The US Army Corps operated additional gauges within the system, but only one gauge record from near Portland (Kelly Point, 1901-1914) has been found (Hickson, 1912; Talke & Jay, 2017). After the start of the Tongue Point tide record in 1925, an additional tide gauge was installed at the Pacific Power and Light power plant in Youngs Bay from March 1931 to February 1943 (see Figure 1). NOAA also acquired nearly 2 years of data at Fort Stevens from Nov. 1940 to July 1942 (NOAA Station 9439008), part of a larger effort to gauge the Columbia River (~18 gauges total). A gauge was placed at Hammond (Station 9430911), within 1km of Fort Stevens, between July 1983-January 1989, and from July 2011 through July 2014. Additional short time series are represented on the summary sheets available at NOAA (see Supplement section S.1). A number of the 20th century tide archival records and associated metadata are available for download at the EV2 database at the National Centers for Environmental Information (https://www.ncdc.noaa.gov/climate-information/research-programs/climate-database-modernization-program). To quality assure 20th century records, we downloaded and digitized manual measurements of water levels made on the tide staff at the Tongue Point gauge from 1925-1957. Staff measurements were made nearly once a day (20-30 times a month), and tabulated along with a simultaneous measurement from the automatic gauge. The staff/gauge comparisons enable an assessment of data quality. See Table 1 and Supplement section S.1 and S.2 for details on data collected and digitized.
2.1.3 Astoria radar tide gauge measurements, 2015-present

In October, 2015 we installed a Campbell Scientific radar tide gauge (model CS476) at the River Pilots Dock in Astoria, Oregon, within 100m of the location of the 1853 gauge (Figure 1; see supplement section S.2.10 and supplemental Figure S.2.31 for more information). The radar gauge zero was surveyed relative to local benchmarks with a 0.001m accuracy in May 2016. A leveling survey in August 2019 confirmed the datum to within measurement precision (0.005m). Measurements continue to the present at 1 minute intervals, with a manufacturer-reported accuracy of 3mm for each individual measurement.

2.2 Digitization

The partial record of water level tabulations that we found for the Astoria station between 1853 to 1876 was digitized into spreadsheets (see Table 1). Data were quality assured for typographical errors and spurious data, using the techniques discussed in Talke and Jay (2017). Extensive portions of the record were missing. Hourly records only covered the 1870-1876 time period, and were incomplete for the year 1870. High/Low tabulations were originally made for the entire 1853-1876 record, but records for 1856, 1857, and 1859 were missing. Similarly, tabulations of the 19th century staff/gauge comparisons were not found for part of 1859 and 1868-1870.

Because the 19th century hourly and high/low record was incomplete, we took pictures of 219 marigrams (each covering a month of data) from 1853-1875. Each marigram (see Supplement Figure S.1.9) is a ~20m by 0.35 m scroll of paper that was stretched over rollers and moved forward by a clock mechanism. As tides rose and fall, water level changed in a stilling well and moved a float up and down. A chain attached to the float subsequently moved a pencil up and down on the paper, producing a continuous pencil trace (Figure 2; see Pugh & Woodworth...
(2014) for more information on float gauges). Once or twice a day, the tide observer tabulated the height of the automatic gauge relative to a nearby tide staff, in a document that was sent monthly to Washington DC via steamship from San Francisco. Each staff/gauge comparison was additionally marked on the marigram by a (now faint) vertical line (Figure 2). Additionally, the observer noted the date, time, water temperature, and staff/gauge difference (see handwriting in Figure 2). The vertical height of the staff zero was defined by leveling the 14.5 foot mark of the staff to nearby benchmarks (see supplement sections S.1 and S.2 for more information).

In total, we took approximately 14,500 pictures of marigrams (Fig. 2) at the US National Archives in Kansas City, Missouri, using a 16.8 Mega-Pixel Nikon D5100 camera. About 60-70 pictures were taken for each monthly tide roll. The camera was mounted on a camera stand to obtain clear and consistent pictures. To avoid data loss, an overlap of ~50% with each previous image was used. A 50mm low-light lens with low barrel distortion (F-stop 1.8) helped minimize camera-based artifacts and enabled sharp contrast between the pencil line and the background paper. Any residual image distortion was rectified in post-processing (see supplement section S.3 for more details). The paper used before April 1861 was darker, grainier, and of poorer quality, reducing contrast. Data was missing for all of 1854, Nov. 1858, May 1862, and May 1868. Only 25 of 70 marigrams between 1871 to 1876 were photographed, and none digitized, because a complete hourly data set was available from 1871-1876.

The tidal trace in each image (e.g., Figure 2) was found digitally by using two line-finding algorithm, which were then compared for consistency. One algorithm used the contrast between dark and light pixels to estimate the line. The other approach sequentially added points to a user-defined starting (seed) point using a nearest neighbor approach, under the criterion that points must have a similar (dark) color. After an initial estimation of the line, outliers were culled, the search area was further restricted, and the algorithm repeated until convergence on one line was obtained. Occasionally, manual adjustment and culling of bad data was necessary when the algorithm worked imperfectly.
The time and height coordinates of the digitized tidal trace was found by first defining known
time and height coordinates on each image. Known points included the hourly time coordinates
written on each marigram (see bottom of Figure 2), the time and height of daily staff
measurements (intersection of vertical line and tidal trace in Figure 2), and the time and height of
high and low waters, which were denoted by a penciled-in circle (see Figure 2 and Talke & Jay,
2013). A linear transform was defined between the pixel coordinates of the image and the
available time and height coordinates. This yielded an estimate of the tidal trace at a higher than
once-a-minute frequency. However, the effective resolution was likely less (order of 5-10
minutes), because the use of a stilling well (as done historically) effectively increases the
response time to water level perturbations (e.g., Agnew, 1986). Here, we focus on hourly records
which were produced by taking the median value of estimates within 3 minutes of each hourly
ordinate. More details about our digitization are given in supplement section S.3.

### 2.3 Leveling History and Datum Reconstruction

To evaluate water level trends for the entire 1853-1876 data set, we (a) reconstruct a detailed
gauge history between 1853 and 1876 using letters, notes, and other metadata (Table 1); (b)
document the history of leveling surveys and the MLLW datum between 1853 and 1925 and (c)
analyze the connection between a 19th century benchmark, USE A-1 (destroyed in 1931) and
pre-1931 benchmarks that still exist.

#### 2.3.1 Astoria Benchmark History

The first tidal benchmark (BM #1) was inscribed on a large rock (the “tide rock”) at the
historical shoreline on July 13, 1853, three days after installation of the tide gauge. A second
tidal benchmark (BM #2) was added in 1862, a third in 1876, and 4 more in 1887 (see
Supplement S2.3). Other early benchmarks included U31, a benchmark used by the town of
Astoria (first verified reference: 1876) and USE A-1 (also known as A32), a US Army Corps of
Engineers benchmark already described as an ‘old’ benchmark by Hickson (1912). With the
possible exception of BM #1 and BM#2, none of the nine 19th century benchmarks are known to
have survived past 1931; moreover, BM #1 and BM#2, if they still exist, are inaccessible (see
Supplement S2.3 and S2.4 for details on BM history). Since there is substantial vertical land
motion in the estuary (particularly near the coast; see Results), benchmarks are moving relative
to a fixed, geocentric datum such as NAVD-88. Hence, an inferential approach that takes into
account local patterns of vertical benchmark stability and vertical land motion is required to
make an estimate of the change in water levels since the 19th century (see also Results).

Seven still-extant benchmarks in Astoria were established between 1920 and 1931 and were
surveyed relative to USE A-1 (Avers, 1926 and Rappleye 1932). Several (e.g., T100 and Y100)
are located on piers over former mudflats and are probably unstable; a third, F31, is documented
to be unstable by the National Geodetic Survey and is near a known landslide (Burns and
Mickelson, 2013). Based on these considerations, we infer that the remaining cluster of 4
relatively stable, pre-1931 BMs best approximates a locally stable datum (this datum moves
vertically relative to NAVD-88, however; see Results). Through these benchmarks, we can
obtain estimates of the 19th century datum. See Supplement Section S.2.3.6 for more
information.

2.3.2 Leveling, Tide Staff and Gauge History, 1853-1876

Documentary evidence found in archival letters and notes reveals that 11 leveling surveys were
conducted between four separate tide staffs and Benchmark #1 between 1853 and 1876 (see
Table S.2.2 in supplement section S.2.4 for detailed information). The 1876 level loop was run
twice, to an accuracy of better than 0.01m. An 1887 survey reported an accuracy of 0.008m. The
accuracy of earlier surveys is not known, but is assumed to be of similar precision for surveys
from 1858 to 1872, given oversight by the same tide observer (Supplement Section 2.3).

Our reduction of data to a stable staff zero suggests that datum accuracy steadily increased over
the course of the 1853-1876 measurement period (Table 2; see Supplement Section S.2.7 for
more details). From 1853 until the gauge was moved in December 1855, notes indicate the staff
was not perfectly vertical and repeated leveling indicated significant subsidence (~0.2m) in the staff zero. The gauge was moved and the staff replaced in December 1855 and January 1856, respectively, at a known vertical offset to the 1853 staff (see supplemental Table S.2.1 in supplement Section S.2.2 for details of staff and gauge changes). Inferential evidence and letters suggests that the staff was stable until late 1858, when repairs to the wharf appear to have caused subsidence of ~0.05m. The post 1858 datum is better constrained with metadata than pre-1858, though a leveling survey in 1862 disagrees with the 1861 and 1867 surveys, probably indicating leveling error but also possibly revealing undocumented shifts in the staff/benchmark relationship. Drifting logs from upstream nearly broke the tide box (stilling well) in 1866 (Supplemental Figure S.4.7), and one of two tide staffs was “carried away” in May 1866 (Supplemental Figure S.4.8). Rotten piles necessitated a new wharf (and staff) in 1867, and siltation of the harbor due to development led to an additional move in 1872.

We deal with the challenges and ambiguities described above by considering diverse interpretations and the influence of each on the datum. Judgement is then used to assess the most likely staff datum scenario, but other scenarios are retained as an envelope of possible uncertainty (see results and notes in Supplement S.2.4 and Table S.2.3). As an example, the 1861 leveling survey is considered more accurate than the 1862 survey because it was made by a trusted source rather than a local unknown (in 1862; see Supplement S.2.4.2). The aggregate result of considering different interpretations is an expanding cone of uncertainty, with the most plausible datum before 1858 significantly more uncertain that the 1870s datum (Table 2; see also Supplement Table S.2.3 and results).

Notes suggest that data quality significantly improved in July 1858 when the original observer (J. Wayne) was replaced by Louis Wilson, and after a new tide stilling well was constructed in August 1858 because 2.75 ft. (0.84 m) of silt had collected in the old. Moreover, the tide gauge house and piers upon which the gauge was mounted were improved in late 1858/early 1859, reducing and almost eliminating accidental gauge stoppages in the marigram record. A new rigor is also evident in the daily staff/gauge comparison, which increased from 10-15 per month to around 60. This intensive attention to the gauge yielded a data set with an extremely consistent staff/gauge offset, which remained a constant 1.89-1.9 feet (0.58m) from 1859 until November
1872, when the offset changed from 1.9 feet to 2.0 feet (1.61m). Interestingly, the 1858-1876
gauge/staff comparisons show less variability than post-1925 Tongue Point data, where
variability between individual gauge/staff checks of ± 0.02m was typically observed (see
Results). Therefore, 19th century tidal records are not necessarily inferior to modern data, at least
up to the recent digital era, post-1990 (see e.g., Ray & Talke, 2019).

2.3.3 Tying Historic to Modern Data

Astoria benchmarks and datum levels were revisited and re-leveled approximately once every
decade during the 1876-1925 period, increasing the likelihood that the datum was carried
forward faithfully. Records or extracts of 4 leveling surveys from three federal agencies (US
Coast and Geodetic Survey, US Army Corps of Engineers, and the USGS) survive and were
recovered (see supplement section S.2.5 for details). From a summary sheet of the US Coast and
Geodetic Survey made in 1924 (see supplement Figure S.2.1), the vertical height of the USE A-1
benchmark referenced in Hickson (1912) is defined relative to BM #1 from 1853. In turn, the
height of benchmark USE A-1 was defined relative to four stable, and still extant, Astoria
benchmarks by Rappleye (1932). From these four benchmarks we obtained 4 estimates of
benchmark BM#1 in the NAVD-88 reference frame. A fifth estimate of BM#1 height in NAVD-
88 was obtained by determining the height of benchmark USE A-1 relative to the staff zero of
the 1925 Tongue Point gauge. The connection was determined using the relative height of USE
A-1 and Tongue Point benchmarks found in Rappleye (1932), and annual surveys between the
1925 staff zero and Tongue Point benchmarks (see Burgette et al., 2009 and Supplement Figure
S.2.20). From these 5 ties, our analysis suggests that the 1872-1876 staff zero (datum that we use
for 1853-1876 data) was 1.484 ± 0.02m below the NAVD-88 datum. The uncertainty estimate is
based on the spread from different benchmark ties and known leveling error. More information
about the datum tie and the uncertainty estimate is provided in Supplement S.2.5 and S.2.6.

Consistent with the variability in modern benchmarks, some 19th century benchmarks may have
settled slightly relative to each other. Between 1876 and 1887, BM #2 and BM#3 had possibly
subsided ~0.006 and 0.015 m relative to BM#1. Similarly, the 1898 USGS survey suggests that
BM#4 and BM#5 subsided by ~0.015m relative to BM#2 since a 1887 survey. However, this
small variation in benchmark heights may also simply reflect leveling uncertainty, which was ~0.01 and 0.008 m in the 1876 and 1887 surveys. Given this uncertainty, we therefore assume that BM#1 and USE A-1 reflect approximately the same vertical land motion rates.

2.3.4 Corrections to 1925-1960 datum at Tongue Point

Burgette et al. (2009) determined that the Tongue Point datum was based on an unstable benchmark between 1925 and about 1960, and therefore applied a linear correction to sea-level data. We assess the stability of the datum at a finer resolution by reanalyzing the original data reduction process on a monthly basis. Evaluation of staff/gauge comparison sheets from the EV2 database (see section 2.1) shows that data was reduced each month by (a) transcribing hourly and high/low data from the marigram, in the vertical frame of reference of the automatic gauge; (b) shifting tabulations into the frame of reference of the fixed tide staff, using the average of 20-30 staff/gauge comparisons; and (c) adding an additional offset to reduce data to a “fixed datum.” (See Supplement S.2.6, Agnew, 1986, or Talke et al., 2018 for examples of comparison sheets and more details on processing.) Further, the staff zero was leveled approximately once a year to available benchmarks (see supplement Figure S.2.20, or Burgette et al. 2009). Therefore, the height of the fixed datum relative to contemporary benchmarks is definable. We use this information to evaluate the “fixed datum” relative to Tidal #7, a benchmark established in 1939 and identified by Burgette et al. (2009) as stable. From 1925-1938, the height of Tidal #7 was estimated using the Burgette et al. (2009) regression slope between Tidal #1 and Tidal #7. Our evaluation of staff/gauge comparisons confirms the Burgette et al. (2009) conclusion that the “fixed datum” used for reduction varied substantially before 1960. A vertical offset (error) of 0.05-0.07 m occurred in the original hourly tabulations (EV2 database) from the 1920s and 1930s, relative to the datum used today. The offset decreased to 0.02-0.04 m in the 1940s and <0.01 m after about 1955 (see supplement Figure S.2.28). Early instability in the “fixed datum” (e.g., 1936) correlated with the occasional installation of new tide staffs; however, beginning in the mid-1940s, adjustments in the “offset to a fixed datum” occurred more frequently, for unknown reasons (Supplement Figure S.2.28). Some of these offsets were corrected ex-post-
facto by NOAA in a piecewise (monthly) fashion (as also occurred with the Boston record; see Talke et al., 2018). Hence, the original EV2 hourly tabulations from 1929 to 1943 differ by 0.006-0.05m from the digital values in the modern NOAA database (see supplement Figure S.2.25 and S.2.26). Even after the 1929-1943 correction, the modern NOAA datum still varies from a datum fixed to Tidal #7. The datum offset is ~0.05m in the 1920s, generally trends downwards in an up-and-down stair-case pattern, and become negligible at the end of the 1950s (green line, supplement Figure S.2.28). As inferred by Burgette et al. (2009), the error (after correction) primarily occurs because Tidal #1 was used as the primary benchmark from 1925-1960, even though it was later shown to be unstable. Our analysis shows, therefore, that the original reduction, plus the later revision, requires a piecewise linear correction pattern to obtain a stable datum (green line, Figure S.2.28).

After correcting for the drift of the “fixed datum” used for reduction, we find that the revised frame of reference is relatively stable compared to benchmarks in the city of Astoria. For example, the vertical drift in the relative heights of benchmark X100 and the revised Tongue Point datum is approximately 0.024m since 1930 (see supplement for benchmark locations). This close agreement suggests that any relative vertical motion between our corrected Tongue Point station datum and a (reasonably) stable benchmark in Astoria is small.

2.4 Tidal analysis

As shown in Jay (2009), the M2 and K1 tidal constituents as Astoria Tongue-Point (Figure 1) increased at a rate of 77 ± 7 and 35 ± 4 mm/century between 1925 and 2007. We determine whether these trends extend to the mid-19th century by applying tidal harmonic analysis (method of Pawlowicz et al., 2002; Leffler and Jay, 2009) to each year of available hourly records from 1855 to 2018. To enable estimation of river discharge (see below), we also estimate tidal constituents using a sliding window of 32 days of hourly data that is incremented forward daily. To account for the astronomical variation in tides, an admittance between the estimated M2 tidal amplitude and the gravitational potential (see e.g. Cartwright and Edden, 1973) is constructed, following Moftakhari et al (2013), and denoted $|M_2|_{adm}$. 
We also evaluate the change in mean tidal range (MTR) over time, where MTR is defined as the difference between annually averaged mean high water (MHW) and mean low water (MLW). MTR was also determined using a sliding 30 day window, using available hourly records. A bias correction of 0.023 m (modern) and 0.027 m (historical) was added to results to account for the slight underestimate in MTR that occurs when using hourly data, compared to high/low data. The correction was based on times of coincident measurements.

2.5 River Discharge Estimation

Mean water levels (MWLs) in Astoria are influenced by river flow, due to its inland location (Chelton & Davis, 1982; Jay et al., 2015). On a monthly time scale that averages over spring-neap tidal variability (Kukulka & Jay, 2003; Jay et al., 2011), this can be expressed with a hydrologic rating curve (see e.g., Kennedy, 1984):

\[ WL \approx a_0 + b_0 Q_r c_0, \]  

(1)

where \( WL \) is the tidally averaged water level ([m]), \( a_0, b_0, \) and \( c_0 \) are empirically derived regression coefficients, and \( Q_r \) is the river flow ([m\(^3\)/s]). In practice, Burgette et al. (2009) regressed river discharge against measured mean water level at Astoria, after removing oceanic sea level variations. Similarly, we subtract out the monthly oceanic variability measured at Neah Bay (NOAA station 9443090) since 1934 from monthly Astoria sea level. The resulting, ocean-corrected water level (mostly due to upwelling variations) is then bin-averaged by river flow measured at Beaver Army Terminal (Rkm 86) in 25 equal increments from 1000 to 25000 m\(^3\)/s (the bin averaging enables the same statistical weight to be given to high and low flow data; see Moftakhari et al., 2013). A linear regression (\( c_0 = 1 \)) finds that \( a_o \) and \( b_o \) in Equation 1 are -0.09 m and \( 1.57 \times 10^{-5} \) s/m\(^2\) (\( R^2 = 0.98 \), p-value < 10\(^{-9}\); using daily-averaged data, \( R^2 = 0.55 \), indicating that some sources of variability were eliminated by bin-averaging). While no nearby gauge is available to remove oceanic fluctuations in the historic record, we find a similar regression slope of \( b_o = 1.44 \times 10^{-5} \pm 7 \times 10^{-6} \) by regressing the annual maximum of monthly
averaged discharge from 1855-1876 against 30d averaged sea level (see below for method used to estimate discharge). A slight, ~10% reduction in the regression slope is found in 1934-1960 data vs. 1980-2018 data. These considerations suggest that the $Q_r$ vs $WL$ relationship is relatively stable within the estuary, but may shift slightly over time.

We next assess whether the seasonal variation of sea level or sea-level rise has been influenced by changing river discharge (e.g., Naik and Jay, 2011). Due to a lack of discharge records pre-1878, we hindcast river discharge from 1855-1876 using tidal data by following these steps:

1. We develop a rating curve between water level in Portland from 1880-1900 and discharge estimates at the Beaver Army Terminal (Rkm 86). The river discharge at Rkm 86 was estimated for the 1880-1900 period by Naik and Jay (2011) based on Columbia River discharge from The Dalles and Willamette River discharge at Albany (see Figure 1).

2. Using the rating curve from (1), we next estimate river flow during 1876 using measured Portland water levels. We use 1876 because this is the only time period in which Portland and Astoria data coincide in time (see Table 1).

3. Using the observation that tidal properties are affected by river flow (Godin, 1999; Kukulka and Jay, 2003; Moftakhari et al., 2013), we develop a calibration between 1876 discharge and 1876 tidal properties. Fortunately, because 1876 was one of the 5 largest floods in the last 170 years, we are able to calibrate to a large dynamic range of discharge conditions.

4. We hindcast discharge from 1855 to 1876 using the calibration from (3)

Though Portland is on the Willamette River (see Figure 1), backwater from Columbia River snowmelt flows dominate water level during the May to September time period, particularly before 1900. Following Helaire et al. (2019), we use this May-September time frame for calibration (steps 1 to 3 above), because it avoids unsteady water level fluctuations (and therefore scatter) associated with short time period (<1 week), winter-time Willamette River discharge events. We regress 30d average water levels against 30d average Columbia discharge and obtain the following regression curve (see Results):
\[ Q = 1800 + 2500h + 1.04h^4 \]  

(2)

Where \( h \) is the depth of water ([m]) above the USGS gauge zero in Portland (1.53m above NAVD-88 datum) and \( Q \) is river discharge ([m^3/s]). The nonlinear \( h^4 \) term was found by empirical experimentation and arises during overbank flow during high discharge conditions (see results). This rating curve (Equation 2) is then used to hindcast 1876 river discharge based off of Portland water level data.

The method of Moftakhari et al. (2013) is next used to relate tidal properties in Astoria (1855-1876) to river discharge. From time series of the \( M_2 \) admittance (section 2.4), we estimate the \( M_2 \) admittance anomaly \( |M_2|_{adm}' \), defined here as the deviation from baseline, low discharge conditions between mid-August to mid-October (prime denotes an anomaly). Specifically, the admittance anomaly \( |M_2|_{adm}' \) is defined by subtracting a baseline value of 0.82 from the admittance, after first removing a slight trend in the 19th century admittance. The absolute value is taken to retain positive values. We also construct time series of the \( M_4/M_2^2 \) ratio, because this ratio also varies with river discharge due to frictional interaction with tidal currents. The deviation of the overtide ratio \( M_4/M_2^2 \) from low flow conditions, denoted \( \left( \frac{M_4}{M_2^2} \right)' \), is obtained by subtracting out a baseline value of 0.02 from the calculated \( M_4/M_2^2 \) ratio (prime denotes an anomaly).

As shown in Jay and Kukulka (2003), river flow \( Q \) can be related to a tide ratio \( P \) (e.g., the \( M_2 \) admittance or the \( M_4/M_2^2 \) ratio) by a power law. The following basis function is used to relate the \( M_2 \) admittance anomaly \( \left( |M_2|_{adm}' \right) \) to water level in Portland, and through that to river discharge (Equation 2):

\[ h = a_1 + a_2 \sqrt{|M_2|_{adm}'} + a_3 |M_2|_{adm}' \]

(3)

where \( h \) is the 30d moving average of water level in Portland, the \( a_i \) are regression coefficients and are equal to \( a_1 = -18.1, b_2 = 209, \) and \( b_3 = -241, \) and the brackets denote that we have taken the amplitude of the anomaly. The square root term arises because overbank flooding at high
river discharge alters the relationship between river discharge and water level, and is a typical feature of rating curves (Equation 1).

The anomaly of the $M_4$ overtide ratio is found to fit 30d averaged water level in Portland, $h$, as follows:

$$h = b_1 + b_2 \sqrt{\left(\frac{M_4}{M_2^2}\right)'} + b_3 \left(\frac{M_4}{M_2^2}\right)' ,$$  \hspace{1cm} (4)

where $b_i$ are found by linear regression to be $b_1 = -9.27$, $b_2 = 130$, and $b_3 = -31.6$. This calibration is valid for 30d averaged water levels between 1.7 and 8m, the dynamic range of water level in the 1876 Portland data, but leads to unrealistic water level estimates below this range (especially for $(\frac{M_4}{M_2^2})' < 0.005$). To estimate water levels below 1.7m, we instead extrapolate using a linear regression with $b_2$ set to zero. The values of this calibration are $b_1 = -.3752$, $b_2 = 0$, and $b_3 = 427$.

Note that the optimum calibration for Equation (3) and (4) occurs when the Astoria tide property is lagged 3d relative to river discharge. The time lag occurs in part due to the barotropic response time of tides to changing water level (river discharge) in Portland, but also likely because the same river flow produces slightly different water levels during the rising and falling arm of a freshet (e.g., due to the effect of storage in wetlands; see Helaire et al., 2019). We estimated the standard error in our estimates following the method described in Moftakhari et al. (2013).

3.0 Results

We evaluate our digitization below (Section 3.1) and then evaluate changes in tidal properties (Section 3.2), sea level (section 3.3), and river flow (section 3.4).
3.1. Data Assessment

Examples of digitized data from the January 1862 marigram (blue curve) and associated High/Low tabulations (green) and staff readings (red) are shown in Figure 3. At a time resolution of 6 minutes, the infra-gravity waves (30-60 minute period) apparent in the pencil trace from Figure 2 are reproduced (Figure 3a). At the 5-day time scale, the diurnal inequality becomes apparent (Figure 3b), while over the monthly time scale the spring-neap cycle is seen (Figure 3c). Close inspection reveals a few data gaps during the month. These gaps occur occasionally throughout the data set and are caused by gauge malfunction or by problems in the line finding algorithms. Typical issues include indistinct pencil marking, text which obscures the curve (see Figure 2), clock issues, or chain adjustments by the observer which altered the height of the pencil trace. Nonetheless, excluding 3 months for which marigrams were not found, we digitized 97% of the hourly records between January 1855 and December 1870. If one includes the digitized hourly tabulations from 1871-1876, nearly 190,000 hourly records were digitized. Marigrams were not found for 1854, and marigrams from 1853 were of marginal quality and therefore not digitized.

Quality assurance and comparison of the different data sets (Figure 4 and 5) reveals the following:

1. A small average difference of <0.003m is found between marigram-derived estimates (1855-1870) and staff measurements made within 6 minutes of the hour (Figure 4). The standard deviation of this difference decreased from ~0.02m to 0.01m between pre-1860 and post-1860 data, indicating that data quality increased during the 1860s (Figure 4a and 4b).

2. Nineteenth century hourly tabulations for the years 1871-1876 are biased nearly 0.01m lower than staff measurements made within 6 minutes; Figure 4c). This suggests that the scaling used by tabulators (called “computers”) during this period to convert from the tide roll to water level was slightly biased (see section 2 and Supplement S.3). The standard deviation of ~0.02m also indicates that there is slightly more error in their
reduction of hourly data than our software-based digitization from the 1860s, which has a standard deviation of ~0.01 m (compare Figure 4b and 4c).

3. By contrast, 19th century estimates of LW and HW are an average of 0.006 and 0.015 m higher, respectively, than tide staff measurements made within 6 minutes of LW or HW (Figure 5). The discrepancy is larger for high water, again suggesting a small error in the scaling used in the 19th century data reduction.

Our results suggest, therefore, that the manual reduction process (which continued in a similar fashion until the advent of digital data loggers in the late 1960s/early 1970s) constitutes a small (±0.01 m) source of uncertainty in historical compilations of sea level.

The observed variance around the mean in Figures 4 and 5 likely has multiple causes. Because historical high/low and hourly measurements were rounded to the nearest 0.05 ft. (0.015 m) and 0.1 ft. (0.03 m), respectively, comparisons between modern computer-based digitizations and historical tabulations will always show some differences. The staff measurements themselves may have some error and bias. Although staff measurements were recorded to 0.01 ft resolution (3 mm; see supplement S.1, Figure S.1.3), the actual error in the measurement may have been more, particularly during stormy conditions. Similarly, the scaling from paper to time/height can also introduce error (it was not exactly 14:1). In our reduction, variance can also be caused by residual distortion in the images (for example from camera lens imperfections or undulations in the paper) or change to the paper over time (e.g., shrinkage). The historical gauge itself may have had some manufacturing anomalies; for example, a note from the observer states that he modified the rollers to produce more accurate comparisons between the gauge and the staff. Other sources of random error include the line-width of the pencil (small) and infragravity waves at HW or LW, which often led to a turning point estimate that was not actually on the curve (see Figure 2; also Talke & Jay, 2013). Clogging of the stilling well or clock issues can also produce systematic errors (e.g., IOC, 1985; Agnew, 1986; Zaron & Jay, 2014). However, the overall accuracy is quite good.
Many of the intermittent issues and small imperfections noted in the 19th century data set are also present in the modern, pre-digital data sets such as Astoria Tongue Point before ~1990 (see Smith 2002 for a discussion of historical gauge types). An evaluation of gauge checks shows that the recorded difference between the staff measurement and the automatic gauge tended to vary slightly day-to-day (see supplement S.2.6). We quantify the variation by calculating the root-mean-square error (RMSE) for each month of staff/automatic gauge differences that was digitized (Figure 6a). As can be seen, the monthly RMSE between staff and automatic gage measurement typically varied between 0.015 and 0.04 m; the average is 0.023 m. Notes confirm that the gauge was functioning poorly during periods of larger RMSE; for example, a note on April 22, 1952 states that “intake to float well was cleared”. It is also likely that the large flood of May-June 1948 (second largest on record) filled the stilling well with sediment, affecting measurements for the following year.

The variation in the offset between staff and gauge measurements (see supplement section S.2.6.1 and supplement Figures S.2.21 to S.2.24 for discussion and archival examples) is important because data was reduced to the staff datum by taking the average of monthly differences (see also Smith, 2002); hence, for time periods with a large variance, the confidence in the mean sea level goes down. Under the assumption that error is random, the 95% confidence interval in the mean can be approximated as \( t^* \frac{\sigma}{\sqrt{N}} \), where \( N \) is the number of measurements, \( \sigma \) is the standard deviation, and \( t^* \) is the t-score and varies between 2.05 and 2.1 for the 95% confidence based on the 20-30 samples typically available each month. Results show that the 95% confidence in the monthly mean is typically around 0.01 m, and ranged between 0.005 and 0.015 m in approximately 91% of the 421 months evaluated (Figure 6b). Occasionally larger imprecision occurred in 1934 and between 1947 and 1952 (Figure 6b). Uncertainty in the mean slightly increases after 1946 due to a reduction in the number of monthly staff checks from ~30 to ~20, in addition to the periods of lesser data quality referenced above. We find no evidence of systematic bias in monthly or annual averages caused by gauge errors, unlike Agnew (1986) did for Port San Luis, California. However, the estimates of gauge precision in Figure 6 do not address uncertainty due to leveling, benchmark or dock instability, or other causes. To assess
such systematic error, a re-evaluation of how data was reduced to a stable datum, such as described in Section 2.3, is necessary (see also supplement Section S.2).

3.2 Changes to Tides

The mean tidal range (MTR) at Astoria Tongue Point increased by 0.10m between the 1858-1876 period and the 1998-2016 period (Figure 7), from 1.98m to the present-day 2.08m. This increase is driven by an upward trend in the $M_2$ tidal constituent after the start of the modern record (Figure 8). Other major constituents ($O_1$, $K_1$, $S_2$, and $N_2$) also increased significantly after 1925 (Figure 8). No significant difference in annual mean tidal range is observed between the 1853-1876 data set and 1925-1945 Tongue Point data, after adjusting 19$^{th}$ century data upwards by 0.05m to account for the modern difference between these locations (Figure 7a). Hence, construction of the jetties from 1881-1917 and alteration of the estuarine mouth (see Sherwood et al., 1990) resulted in little net change in annual mean tidal range in Astoria. However, the relative magnitude of tidal constituents shifted between 1876 and 1925; a slight increase in $M_2$ occurred, while the $O_1$, $K_1$, $S_2$, and $N_2$ constituents slightly decreased (Figure 8). Some of the observed changes in constituent amplitudes may be attributable to spatial variability. However, most present-day constituent amplitudes at the historical (Astoria) and modern (Tongue Point) measurement sites are quite similar (Figure 8), suggesting small or negligible spatial gradients. The exception is $M_2$, which increases by 0.02m in amplitude in the 5km between Astoria and Tongue Point (Figure 8a), due to a partial reflection at the Tongue Point headland (Giese and Jay, 1989). Given that Astoria and Tongue Point tides are similar, we conclude that observed 20$^{th}$ century trends in $M_2$, $S_2$, $N_2$, $O_1$, and $K_1$ (Jay, 2009) do not extend back into the 19$^{th}$ century.

The observed increase in the $M_2$ tide over the past century is coupled with a decrease in the $M_4$ overtide (Figure 8). At Astoria, $M_2$ increased by ~5% between the 1855-1876 period and our recent measurements from 2015-2019 (Figure 8). Similarly, $M_2$ at Tongue Point increased by ~7% (Figure 8; also Jay, 2009). As $M_2$ increased, the $M_4$ overtide decreased from 0.025-0.04m in the 19$^{th}$ century to within 0.003-0.015m in recent years. Because $M_4$ overtide production within a
large river is primarily caused by non-linear frictional interaction between the M\textsubscript{2} tide and river flow (Godin, 1999; Kukulka & Jay, 2003), a reduction in M\textsubscript{4} coupled with an increase in M\textsubscript{2} suggests that frictional drag in the estuary has decreased since the 19\textsuperscript{th} century.

Anthropogenic modifications over the past 150 years to the LCRE (e.g., Sherwood et al., 1990) are likely a primary cause of the observed trends in tides. A depth-averaged model recently showed that a doubling of the shipping channel depth and other bathymetric changes since the late 1800s increased M\textsubscript{2} amplitudes throughout the lower Columbia River, with a maximum about 60km from the coast (Helaire et al., 2019). Tidal change was linked to the decreased frictional damping caused by increased depth. The observed secular increase in M\textsubscript{2} at Astoria could also be influenced by the partial reflection in the M\textsubscript{2} at Tongue Point. In other estuaries, the maximum change in tidal statistics due to channel deepening is often observed at a point of full reflection (Winterwerp et al., 2013; Talke & Jay, 2020). Similarly, the secular change at the point of partial reflection (7\% at Tongue Point) is larger than the change 5km away (5\% Astoria; Figure 8a), suggesting a similar mechanism. Determining the relative influence of partial reflection and channel deepening on secular change is left for future, detailed modeling experiments.

Tides are also likely influenced by the altered seasonal distribution of freshwater discharge and the long-term decrease in annual discharge volume (e.g., Jalon-Rojas et al., 2018; see section 3.3 below). Here, we note first that the magnitude and timing of the seasonal fluctuation in tidal range has shifted over time. Historically, the minimum, 30 day averaged tidal range occurred in mid-June, and the maximum occurred in mid-September (Figure 7b). For the modern period (2000-2018), the maximum 30d averaged tidal range occurs at the end of May/beginning of June, and the annual maximum is closer to October 1\textsuperscript{st}. A secondary maximum occurs at the beginning of March, but did not occur historically (Figure 7b). Taking the difference between the historical and modern curves (Figure 7b), we find that the largest increase in 30 day averaged tidal range occurred in June and July (+0.17m) and the minimum change occurred in winter and early spring (+ 0.07m; Figure 7c). A changed seasonal cycle in tidal range is unlikely to be driven by
bathymetric alteration, but could be caused by river flow changes (see below). Again, detailed 3D modeling would be required to fully assess the reasons for long-term shifts in estuarine tidal properties.

### 3.3 River Flow Changes

The large snow-melt floods of the 19th century produced a substantial effect on tides in Astoria, as is demonstrated through evaluation of the 1876 flood, one of the 5 largest flood events since 1850 (Naik & Jay, 2011; see Figure 9). As water levels in Portland increased, the M₂ tidal admittance at Astoria decreased (Figure 9a). At the peak Portland water level (i.e., peak flood) in mid-June, the M₂ admittance was ~15% less than its late summer baseline, and corresponded to a nearly 0.3m decrease in tidal range. Variations in M₂ admittance anomaly (Equation 3) and \(\left(\frac{M_A}{M_2}\right)\) (Equation 4) also correlate well with Portland water level during 1876 (Figure 9c and 9d). Water level in Portland during late spring and summer was historically driven by backwater from the Columbia River (e.g., Helaire et al., 2019). Therefore, a good correlation is also observed between Portland water levels from 1880-1900 and corresponding estimates of river discharge at the Beaver Army Terminal (from Naik & Jay, 2011) during the May-September period (Figure 9b; Equation 2). Hence, tidal properties are directly related to river discharge magnitudes. (see also Moftakhari et al. 2013, 2016).

We next construct a composite water level data set for Portland from the regression fits shown in Figure 9, using tidal statistics obtained from the 1855-1876 hourly record. The M₂ admittance anomaly (Figure 9c) is used to estimate elevated discharge conditions (water levels > 4.5m), because of its superior fit within this range (Figure 9c; see also Moftakhari et al., 2013). The \(\left(\frac{M_A}{M_2}\right)\) anomaly, \(\left(\frac{M_A}{M_2}\right)\), is used for lower flow conditions (<3m; Figure 9d). A linear combination is used for values between 3 and 4.5m. A comparison of stage estimates against measurements in Vancouver (1872-1876) and Portland (1876) suggests excellent agreement, with a root-mean-square error (RMSE) of 0.032m and 0.046m, respectively (Figure 10a). The slightly larger error
for Portland water level occurs because the tidal-discharge estimation method works best during high flow conditions and the slowly varying spring freshet (the only data available for Vancouver). The method works slightly less well when applied to the short time scale variations that occur in the Portland record during winter-time, lower flow conditions (see also Moftakhari et al., 2013). The hindcast stage estimates from 1855-1876 (e.g., Figure 10a) are then used to estimate discharge, using Equation 2 (Figure 10b). Finally, a hydrograph consisting of a 30d moving average of discharge over the 1858-1876 time period is calculated (Figure 11a).

The 19th century hydrograph shows a distinct seasonal rise and fall during the April to August time period, with a peak in June (Figure 10b and Figure 11a). By contrast, the modern hydrograph is maximum in late May/early June, with a magnitude that is approximately half (54%) of the historic peak. During winter (Dec-Feb), modern discharge is 25-50% larger than historic conditions (Figure 11a). Averaged over 19 years, the annually averaged discharge was 19% higher historically than the 1998-2016 period (Figure 11a). Compared to the discharge of 7,100 m$^3$/s in the 1970-1999 period (Naik & Jay, 2011), the annual discharge from 1858-1876 was ~10% larger. These estimates agree well with the Naik and Jay (2011) estimates of a 17% decrease in discharge in the last century and a 40% reduction in the spring freshet, based on post 1878 discharge data. Our results are therefore consistent with previous observations of a long term decrease in river flow and a shift in seasonal timing. This validates our approach. The changed hydrograph likely impacts the seasonal cycle of water level in Astoria. Using the discharge ($Q_r$) vs. water level ($WL$) relationship discussed in Section 2.4, we estimate an ~0.03m increase in WL in winter, and a nearly 0.15m decrease in July (Figure 11b and 11c); the validity of these estimates is checked in section 3.4 below.

The long-term changes in river discharge explain some of the observed trends in tidal properties. Using the modeled relationship between river discharge and $M_2$ (Figure 9), we estimate that the about 10% of the long-term change in annual $M_2$ (Figure 8a) is attributable to the approximately 1,200 m$^3$/s decrease in annual discharge (Figure 11). Further, the variability in the nodally-corrected, annual $M_2$ amplitude between 0.88 to 0.91m that occurred between 1855-1876 (Figure
8a is primarily forced by variations in annual averaged discharge, which ranged from 4,000 m$^3$/s (October 1868 to September 1869) to 10,000 m$^3$/s (October 1875 to September 1876). Moreover, seasonal changes to 30d averaged tidal range ($\Delta MTR$; Figure 7c) are almost entirely explained by changes in the timing and magnitude of river flow (Figure 11a). Hence, the largest change in MTR (Figure 7c) is observed when the historical/modern change in river flow is maximum, which is observed to occur on the falling limb of the annual spring freshet in June-July (Figure 11a and Figure 11c). Similarly, larger wintertime flows produce a significant reduction in the modern tidal range between early autumn (Sept/October) and December-February; a much less pronounced decrease occurred historically, due to a smaller difference in discharge. A small peak in $\Delta MTR$ occurs in mid-March, potentially reflecting changes in the spring freshet of coastal tributaries such as the Willamette River (see also Helaire et al. 2019).

3.4 Seasonal and Spatial variability in mean water level

The mean water level throughout the estuary varies due to the tidally-averaged slope that drives the mean river outflow (e.g. Helaire et al., 2019). Hence, the elevation of mean water level varies between stations (Figure 12b and 12c). Over nearly four years of measurements, water levels in the city of Astoria (Rkm-24) were an average of 0.015m lower than measurements at Tongue Point (Rkm-29), yielding a slope of 3×10$^{-6}$ (Figure 12c). Similarly, sea level at the Youngs Bay station was 0.057m less than the datum-corrected Tongue Point RSL from 1931-1942 (Figure 12b). Because the entrance to Youngs Bay is approximately 9km downstream of Tongue Point, this yields a slope estimate of 6.3×10$^{-6}$ for this time period, under the assumption that the relatively small Youngs River does not produce an appreciable slope. Finally, a 0.13m mean difference is observed between Tongue Point (Rkm 29) and Hammond (Rkm 11) during the 2011-2014 period, for a mean slope of 7.7×10$^{-6}$. The general agreement in slope is encouraging, and only possible because of more than a century of careful tide measurements and leveling surveys. Nonetheless, because the measurements all span different time periods, the results may be influenced by changes in river flow, bed roughness, channel depth, and other factors such as wind forcing. The benchmarks used to tie to the NAVD-88 datum also exhibit differential rates of vertical motion (Burgette et al., 2009), which may produce uncertainty in the elevation
difference and estimated river slope (see also Hudson et al., 2017). We find a slight, but statistically significant, subsidence of $0.5 \pm 0.3$ mm/year in the Tongue Point gauge, relative to our Astoria gauge (95% confidence, based on monthly differences; $N=45$). This relative vertical motion is consistent with the Burgette et al. (2009) estimate of an $\sim 0.4$mm/year difference in vertical motion between the Z100 benchmark (closest benchmark to our Astoria gauge) and the Tidal #7 benchmark (Tongue Point). Therefore, our estimate of river slope may be affected by differential vertical land motion, and a high precision leveling survey between the tide stations is likely needed to validate our estimate of river slope.

On the century-time scale, changed river flow hydraulics may also affect the mean offset between gauges. Historically, the South Channel (Oregon side of estuary) was much shallower (Helaire et al., 2019) and a larger proportion of flow may have exited through the North Channel (north side of the estuary) than does today (see e.g. Buschman et al. 2009 for discussion of how changes to depth alters flow partition between distributary channels). We conclude, as do Burgette et al., (2009) and Hudson et al. (2017), that inferring mean slopes in an estuary to more than one significant figure remains challenging with available data due instabilities in benchmarks and uncertainty in the underlying datum or geoid (see also Supplement S.2.6). Nevertheless, the inferred mean slopes are of similar order-of-magnitude at all stations.

Comparison of the historic and modern periods suggests that seasonal sea level variations caused by river flow have greatly changed (Figure 13). After removing the effect of river flow (Figure 11b), the remaining seasonal variability (produced primarily by upwelling and downwelling) is similar (Figure 13b). In both time periods, relative sea level is largest in December-January during the period of down-welling winds, and lowest during the summer upwelling period (Figure 13a; see also Strub et al., 1987 and Chelton & Davis, 1982 for causes of sea-level variability). A secondary peak is observed during spring time in both records (Figure 13a), and is caused by the annual snow-melt driven freshet (Figure 11; see also Burgette et al., 2009). Historically, sea level remained elevated from early May through mid-August, reaching a local maxima in mid-June that was $\sim 0.1m$ higher than water levels in April and $0.15m$ higher than the late summer minimum (Figure 13a). By contrast, the local peak in the modern series occurs in late May/early June, with effects that extend over a shorter time-period (early May to end of...
June; Figure 13a). The average spring-time rise in water level to a peak around June 1st was less than 0.03m in the 1998-2016 period. The decreased springtime rise in water level reflects the large decrease in river discharge during springtime (see Section 3.3 and Naik & Jay, 2005 and 2011). Less obviously, the difference between the 25th and 75th percentile water level has decreased by ~20% during this period (fill plot in Figure 13a). This reflects a decrease in the variance in spring-time river flow, in addition to a reduction in the mean. Therefore, flow regulation has prevented both large spring freshets (e.g., 1876; Figure 9) and extreme low flows (Naik et al., 2011), and has tightened the distribution around the mean.

After correcting for river flow effects, the seasonal cycle in water level in both historical and modern periods is similar (Figure 13b). Because wintertime river flows are larger in the modern period, applying the flow correction narrows the difference between historic/modern water levels in the October to March time frame (compare Figure 13a and 13b). The correction also removes the spring freshet, such that the minimum water level in both series occurs in mid-July (Figure 13b). The offset between the two curves (Figure 13b) varies slightly over the year, possibly due to other sources of variability. Nonetheless, the good correspondence between historic and modern water levels is consistent with the interpretation that long-term changes to river flow drive changes to the seasonal mean water level cycle. There is no statistically-significant evidence of changes to summertime upwelling or winter downwelling, though this comparison is unlikely to capture small changes to the timing and duration of these events, and compensating changes in upwelling/downwelling forcing and the estuarine response cannot be excluded. On an annual scale, the overall effect of altered river flow on sea level is minor. Using the rating curve for Astoria (See Section 2.4), we estimate that the larger annually averaged discharge from 1858-1876 raised water level ~0.01m higher than comparable conditions over 1998-2016.

3.5 Sea Level
The exhaustive search of archival data and meta-data described in Section 2 has resulted in a data set that is reduced to a common datum, allowing an analysis of interannual variability (section 3.5.1) and trends (section 3.5.2) in relative sea level (RSL).

3.5.1 Interannual Variability

Nineteenth century sea level in both Astoria and San Francisco shows interannual variability in magnitude of >0.1m, but no statistically significant trend (Figure 12a). In both records, periods of depressed RSL in 1859-1864 and 1870-1875 are bracketed by elevated RSL levels, indicating general agreement. Year-to-year variability is less well correlated between the San Francisco and Astoria records, in part due to different discharge variability in the two systems. Peaks in Astoria MSL in 1866 and 1876 are related to large discharge years (see e.g., Figure 10), and adjusting for discharge removes much of the 1876 anomaly (Figure 12a). Similarly, the ~0.05m rise in 1862 San Francisco sea level over 1861 levels is likely related to the anomalously large river discharge that occurred in 1862 (see Moftakhari et al., 2013).

Large interannual variability in RSL in the lower Columbia River Estuary is apparent in both historic and modern periods (Figure 12 and Figure 14), with the difference between minimum and maximum equal to 0.12 and 0.14m in the 1853-1876 and 1995-2018 periods, respectively, after removing any trend. The corresponding standard deviation is 0.036 and 0.038m (N=24). The consistency between different periods suggests that interannual variability (after correcting for river discharge) has not measurably changed. Such interannual variability increases uncertainty in long-term trend estimates (e.g., Dangendorf et al., 2013) and helps explain why no statistically significant acceleration in sea-level rise since the 19th century is observed in Astoria.

Oceanic processes such as the El Niño Southern Oscillation phenomenon (ENSO) drive much of the variability observed in the sea-level record (Figure 12 and Figure 14). For example, the highest modern sea level at Astoria, as at other stations in the Northeast Pacific (see e.g., Smith...
2002), occurred during the 1982-1983 and 1997-1998 El Niño events (Figure 14). The strong El Niño event of 1940-1942 (Kaplan et al. 1998) and the El Niño event of 1958 also produced peaks in sea level (Zervas, 2009).

Historically, the persistent high sea level from 1865-1869 and the flow-adjusted peaks in sea level in 1866 and 1869 likely correspond to peaks in the El Niño 3.4 index (Kaplan et al., 1998), which is commonly used to define El Niño events. Similarly, the depressed sea level in the late 1850s/early 1860s and the early 1870s corresponds to a persistent period of La Niña conditions (see Figure 12 and 14). While the ENSO event that peaked in 1877 was one of the largest on record (Kaplan al. 1998), its beginning in late 1876 was too late to be captured by Astoria data. On the other hand, the peak in sea level observed at both San Francisco and Astoria in 1855 and the elevated Astoria sea level from 1855-1858 are suggestive of El Niño conditions. Historical reconstructions and proxy analysis (e.g., Quinn et al., 1987; Gergis and Fowler, 2009) suggest intermittent El Niño conditions in the 1853-1858 periods, while the beginning of the El Niño 3.4 index in 1856 is strongly positive (Kaplan et al., 1998). However, since pre-1860 sea level data are less reliable (see Smith, 2002 regarding San Francisco), and because pre-1870 ENSO indices are less certain and sometimes contradictory (compare Kaplan et al., 1998 and Gergis & Fowler, 2009), we caution that these results are suggestive rather than definitive.

3.5.2 Sea Level Rise

After correcting for river flow effects (see Section 3.3 and Figure 11) and the offset due to the hydrodynamic slope between Astoria and Tongue Point (0.015m added to 1853-1876 data; see Section 3.4 & Figure 12c), we estimate that relative (local) sea level at Astoria Tongue Point increased by 0.06 ± 0.04m between the 1858-1876 and 2000-2018 periods (Figure 14a, Table 3). An average river flow correction of 0.011m was subtracted from 1853-1876 data, and an average of 0.007m added to 2000-2018 measurements, reflecting river discharge that was either above or below the long-term average, respectively. Note that the same RSL rise is found using the full 1853-1876 data set, but with larger uncertainty due to a less certain datum in the early 1850s.
By contrast, the relative sea level in the combined Fort Stevens/Hammond record (Rkm 11) decreased on the order of 0.1 m between the early 1900s and the present (Figure 14b, Table 3). Based on a regression of RSL during coincident years, the relative difference in uplift rates between Tongue Point and Fort Stevens is $0.91 \pm 0.45$ mm/yr (N=16). Similar vertical rates are found using GNSS from Tongue Point (station TPW2) and Fort Stevens (stations FTS5 and FTS6), and help confirm our analysis (see Table 3). For example, estimates from SONEL (Système d’Observation du Niveau des Eaux Littorales; Gravelle et al., 2013) suggest that Fort Stevens (station FTS5) is rising at a rate of $0.94 \pm 0.4$ mm/yr compared to Tongue Point (Table 3). Similarly, estimates from the Nevada Geodetic Library updated through August 2019 (see Blewitt et al., 2016 and Blewitt et al., 2018) suggest a relative rate of $1.1 \pm 0.8$ mm/yr (average of the difference between TPW2 and the two GNSS sensors at Fort Stevens; Table 3). While these results are encouraging, we note that an order 1 mm/yr uncertainty is possible in GNSS trends based on the choice of terrestrial reference frame (Wöppelman & Marcos, 2016). Therefore, these results should be revisited in the future as additional records are gathered and the reference frame improved.

Applying the most recently updated values for vertical land motion from the Nevada Geodetic Library (Table 3), we estimate that mean water level at Astoria (relative to a geocentric origin) rose by $0.11 m \pm 0.09 m$ from the 1858-1876 period to the 2000-2018 period (Figure 14c). A rise of $0.11 m$ is approximately half the global rise of $\sim 0.23 \pm 0.025 m$ since the 1860-1880 period estimated by Church and White (2011). Our result is more similar to recent estimates of 20th century sea-level rise, which have been revised downwards to approximately 0.13-0.14 m (Hay et al., 2015; Dangendorf et al., 2017). The lower-than-average geocentric sea-level rise in Astoria occurs in part because of the redistribution of water towards the western Pacific over the past ~30-40 years, which depressed recent rates of sea-level rise in the Eastern Pacific (Merrifield, 2011, Bromirski et al., 2011; Merrifield and Thomson, 2018). Moreover, the lack of historical records pre-1900 makes global estimates uncertain; as indicated by Church and White (2011), their estimate from the 1860s was based on only 7-14 data series that were mostly from the north Atlantic. Finally, our calculation assumes that vertical land motion rates between 1853 to 2000 are the same as that estimated from the GNSS record at Tongue Point over the 2000 to 2019 period (Table 3). The Burgette et al. (2009) evaluation of benchmarks and our evaluation of
relative sea-level rise and GNSS rates at Fort Stevens and Tongue Point are consistent with a constant average rate of VLM; however, time-varying rates of VLM cannot be precluded in a tectonically active region, particularly in the less well-documented period before 1920.

The estimated 0.11m ± 0.09m rise in geocentric sea level at Astoria is larger than, but within the error bounds, of the sea-level rise in San Francisco (Table 3). After accounting for a vertical land motion rate of -1.3 ± 0.6 mm/yr (from station TIBB; see Table 3), geocentric mean sea level at San Francisco rose only 0.024 ± 0.09m from the years 1858-1876 to 2000-2018. Some of the difference in the estimated geocentric sea-level rise (Table 3) for the Astoria and San Francisco stations likely results from uncertainty in the VLM correction. However, GNSS-based estimates of the vertical drift velocity between station TIBB and Astoria Tongue Point are remarkably consistent with our tide-gauge based estimates of 20th century rise (after our correction of the Tongue Point datum). For example, over the 1926-2014 time frame, relative sea-level trends in San Francisco are estimated to be 1.53 ±0.2 mm/yr greater than the datum-corrected Tongue Point record (Table 3). Similarly, GNSS-based estimates suggest that the difference in VLM between San Francisco (station TIBB) and Astoria (station TPW2) is 1.68 ±0.83 mm/yr. The differences in SLR rate between Fort Stevens and Astoria are also similar to GNSS-based rates (Table 3), and the Fort Stevens record collapses onto the Astoria record after applying the vertical land motion correction (Figure 14c).

The good consistency between GNSS and sea-level based estimates of relative VLM increases our confidence in the use of Astoria as a sea-level gauge, and in our early 20th century datum correction. Nonetheless, the good 20th century comparison also suggests that uncertainty in GNSS records is not the only cause of the 0.086m difference in geocentric SLR between San Francisco and Astoria between the 1858-1876 and 2000-2018 periods (Table 3). In both estuaries, some bias may remain due to changed gauge location, altered tidal circulation and decreased river discharge. Within the Lower Columbia River, channel dredging and deepening has reduced the tidally averaged slope and decreased mean water levels, particularly in the tidal river upstream of Astoria (see Jay et al., 2011 and Helaire et al., 2019). Differential land motion adds uncertainty to hydrodynamic slope corrections for gauge location (see discussion above). In addition, changes in estuary geometry have altered circulation patterns (Hamilton, 1990). Within
San Francisco Bay, a 30-35% decrease in annually averaged discharge since the 19th century (Moftakhari et al., 2013, 2015) likely influenced (suppressed) mean water levels. Further, the historical variation in gauge location in San Francisco Bay is hypothesized to cause an order 0.02 m offset in annually-averaged sea level during some periods, based on an approximation of the river slope (Breaker & Ruzmaiken, 2013). Like the lower Columbia River, circulation patterns have also likely changed, as indicated by a ~7% per century increase in the M2 amplitude (Jay, 2009; Woodworth, 2010). Secular trends in tidal amplitudes are caused, in part, by dredging and removal of sand dunes (Rodríguez-Padilla and Ortiz, 2017) and decreased river flow (Moftakhari et al., 2013). As shown in other estuarine systems (e.g., Jay et al., 2011; Ralston et al., 2019), such increases in tidal amplitudes often coincide with a decrease in river slope. Nonetheless, as with the Columbia River, a baroclinic modeling approach is likely required to assess the importance of hydrodynamic changes at a near-coastal station.

Our analysis also supports the Zervas (2009) conclusion that the 19th century San Francisco record may have some small, uncorrected datum shifts that should be further investigated. Periods of dock subsidence occurred in the 1850s and the late 1870s-early 1880s that likely increase uncertainty, even though they were apparently corrected in 19th century tabulations, (Smith, 2002; Talke & Jay, 2013). Further, the San Francisco record is stitched together from data from at least 3 locations (Smith, 1980, 2002; Talke & Jay, 2013); each location shift makes the datum less certain. During the 1877 move from Fort Point to Sausalito (a location about 6 km away), 6 months of simultaneous observations suggest a datum tie that is 0.012 m different than the leveling tie used to connect the series (Zervas, 2009; Smith 1980). Further, Zervas (2009) suggests that pre-1897 sea-level records from San Francisco are biased +0.037 m high due to differential vertical motion caused by the 1906 earthquake. This bias occurs because the Smith (1980) reconstruction of the San Francisco record used post-1906 leveling ties (the only ones available). Using the Zervas (2009) correction, the geocentric rise in San Francisco between the 1858-1876 and 2000-2018 epoch is ~0.06 m, which is more consistent with our estimate of 0.11 m for Astoria. Further work is necessary to reconcile the 19th century estimates at San Francisco with Astoria.
Conclusions

To characterize changes to tides, relative sea level, and river flow since the 19th century in the Lower Columbia River Estuary, we digitized a combined total of ~35 years of tide records from the 1853-1876 and 1931-1942 periods, including hourly, high low, and daily staff measurements. We also digitized approximately 13,500 pictures of marigrams from 1855-1870, yielding a 97% complete hourly record. The historical datum and staff zero was constrained through 11 leveling surveys made between 1853 and 1876, and 4 additional surveys made between 1887 and 1920. The historical staff zero was connected to the modern NAVD-88 datum through multiple benchmarks, yielding both an average tie and an estimate of uncertainty.

Careful evaluation suggests that datum uncertainty and data quality in both the modern and historical data sets varies over time, but is generally small compared to the tidal signal (at hourly resolution) or sea level variability (at monthly or annual time scales). Errors and bias in historical, pre-digital data are caused by multiple factors, including instrumental imprecision (e.g., Figure 6a), the data reduction process (Figures 4-5), and uncertainty in the datum caused by leveling and benchmark instability (see supplement Section S.2.4 to S.2.6). Other factors that cause imprecision include timing errors, infragravity waves, instrument vibration, clogging of the stilling well, and manufacturing inaccuracy. An evaluation of staff, high/low, and hourly measurements from 1855-1876 indicated that further errors were introduced during transcription from the paper marigram to high/low or hourly data. Aggregated over many years of data, a standard deviation of 0.01-0.02m and biases of up to 0.01m were observed between individual 19th century staff measurements and different data types (Figure 4 and 5). A comparison of staff and gauge measurements from 1925-1957 found an overall RMSE of 0.023m, suggesting this order of magnitude precision for an individual measurement. The 95% confidence in measurement precision for a monthly average ranges from 0.007 to 0.015m for most time periods (Figure 6).
The largest uncertainty introduced into the combined 1853-2019 sea level data occurs through leveling uncertainty (particularly during the 1850s) and benchmark instability, leading, for example, to an unstable station datum between 1925 and 1960 at Tongue Point (see also Burgette et al., 2009). While the observed vertical datum drift of ~0.05m may be larger than is typical for coastal stations, a similar problem was observed at Boston (Talke et al., 2018) and at Garibaldi, Oregon (Burgette, et al., 2009). These observations highlight the need to reanalyze existing 20th century leveling records, in addition to data rescue efforts (Bradshaw et al., 2015). A more granular accounting of uncertainty in existing records can lead to improved reconstructions of sea level, particularly methods that require a ‘prior’ estimate of uncertainty (e.g., Hay et al., 2015).

We estimate that a geocentric rise of \( 0.11 \pm 0.09 \text{m} \) (95% confidence) occurred within the LCRE over the last 150 years, after accounting for hydrodynamic effects and vertical land motion. The corrections applied for changed gauge location (+0.015m) and decreased river discharge (-0.01m) are small and partially offset each other. Hence, relative sea level changes in the Lower Columbia River Estuary appear to be driven by sea-level rise in the Eastern Pacific and vertical land motion. Because vertical land motion caused a spatial variation in relative sea-level rise of \( \approx 10\text{cm}/\text{century} \) at two locations within 15km of each other, there is a need for an observation system that can better resolve the spatial gradients in sea-level rise at the local scale, and reduce uncertainty in land motion estimates. Nonetheless, our results are broadly compatible with an estimated geocentric sea-level rise of \( 0.06 \pm 0.09 \text{m} \) in San Francisco (after the Zervas, 2009 datum correction), the only comparable record currently available for the Eastern Pacific.

The high-frequency data digitized from 1853-1876 also enables a long-time scale view of system change, as reflected in tidal dynamics and river flow. A reconstruction of pre-1900 Columbia River discharge suggests that spring freshets were \(~19\%\) larger between 1858-1876 than 1998-2016, consistent with the \( 17\% \) overall reduction in mean flows between 1878 and 2000 found by Naik and Jay (2011). Spring freshets were \(~50\%\) larger historically, and winter flows 25-50\% less. At the city of Astoria, the \( M_2 \) constituent increased by 5\% between the mid-19th century
and the present, slightly less than the 7% change observed at Tongue Point from 1925-present (Figure 8a; see also Jay 2009). The seasonal cycle of 30 day averaged tidal range has shifted, with the largest increases occurring during May-July (Figure 7) due to the large reduction in spring-time flows (Figure 11). The smallest change in 30 day averaged tidal range is observed in winter, because flow has increased and damps the M$_2$ constituent. Overall, the decreased river flow accounts for approximately 10% of changes to the M$_2$ tide. The primary cause is probably other anthropogenic changes such as channel deepening, land reclamation, and inlet alterations, as suggested by Helaire et al. (2019).

The observed changes to tides and river flow observed in Astoria are correlated with hydrodynamically-induced changes to mean-water level from seasonal to secular time scales. The altered seasonal pattern of tidal range and river flow between the mid-1800s and today is correlated with an altered seasonal pattern in relative sea level (Figure 7, 11, and 13). Similarly, year-to-year variations in tidal constituent amplitudes (e.g., Figure 8) caused by river discharge are correlated with variations in annually-averaged water level (Figure 12; see discussion in section 3.3). The evidence presented here suggests, therefore, that changes to tidal constituents within an estuary can be an indicator that the hydrodynamic component of mean-water level has also changed (see also Jay et al., 2011, for the link between mean water level and tidal statistics in a tidal river). Many sea-level stations in global repositories (e.g., Holgate et al., 2013) are located on or near major rivers (e.g., San Francisco, California and Vancouver, Canada) or are located further up a river estuary than Astoria (e.g., Washington DC and Wilmington, North Carolina). These locations are also marked by significant secular trends in tidal amplitudes (Jay, 2009; Woodworth 2010; Talke & Jay, 2020; Haigh et al., 2020), caused for example by channel dredging (Familkhalili & Talke, 2016). Hence, we suggest that the combination of approaches shown here—an evaluation of seasonal, interannual, and secular trends in river flow, tides, and relative sea level—could be used to help determine to what extent other sea-level records within estuaries and harbors are affected by local hydrodynamic changes.

Acknowledgements and Data
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References


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Figure Captions

Figure 1: Site Map of the Lower Columbia River Estuary. Inset triangles show locations of
historical tide measurements (Astoria Tongue Point, Astoria, Astoria Youngs Bay, Hammond,
and Fort Stevens), historical water level measurements (Portland, Vancouver) and discharge
estimates (Beaver Army Terminal; from Naik & Jay, 2011). Our radar gauge is in Astoria, within
100m of the historical gauge location; see supplement Figures S.2.32 and S.2.33 for a close-up
map of its location. The head of tides is at Bonneville Dam, and is located 231km from the River
to km zero (Rkm 0). River km zero is shown with a magenta x, and is located at the intersection of
the shipping channel and a line connecting the seaward end of the North and South Jetties. Place
names are given the same color as water level results from those locations (see Results). The
Columbia River watershed (inset at left) covers portions of seven states and Canada.

Figure 2: Example of a marigram from Astoria, OR from Jan. 1862. The vertical lines associated
with the vertical text denote times at which the observer compared the measurement of the
automatic gauge with a nearby tide staff. Hourly times are tabulated on the bottom of the page,
and corresponded with holes punched into the paper by the roller mechanism which moved
paper. Each horizontal line was an increment of roughly 0.025m, and the pencil trace
downscaled the actual vertical tide by a ratio of approximately 14:1. Infragravity waves are
apparent on the pencil trace, and probably occurred due to storm-conditions in the ocean. This
marigram was taken at the beginning of the most extreme and extended period of cold weather
since ca 1840 in the Lower Columbia River Estuary, and about a month after a large rain-on-
snow event in the Willamette River (e.g., Miller 1999).

Figure 3: Examples of digitized water level data from January 1862 over (a) part of the marigram
image depicted in Figure 2; (b) a 5-day period and (c) the entire month. Digitized tabulations of
high and low water are given by squares, and staff measurements by diamonds. In (a), 6-minute
resolution data (cyan) are overlaid by data with an approximately 1 minute resolution (blue).
This comparison shows how high frequency variation in the approximately 1-minute resolution estimate (see also Figure 2) was smoothed. Hourly resolution is plotted in (b) and (c). Time is in UTC, and the datum is the staff zero. Comparisons between the digitized marigram and available High/Low and staff measurements demonstrate a good fit to data.

Figure 4: Distribution of the difference between hourly data and staff readings from three periods: (a) 1855-1869, (b) 1860-1867, and (c) 1871-1876. Only staff readings within 6 minutes of the hour were differenced. The y-axis depicts the total number of measurements (counts) within each bin. The 1855-1867 distributions are based on our digitization of marigrams, while hourly tabulations from the 19th century were used for 1871-1876. Staff readings were not found for the 1867-1870 time period. The average difference (μ) and standard deviation (σ) of each distribution is given.

Figure 5: The distribution of the difference between (a) tabulated High Water and staff readings and (b) tabulated Low Water and staff readings for the time period 1860 to 1876. The average difference (μ) and standard deviation (σ) of each distribution is given. The y-axis depicts the total number of measurements (counts) within each bin. Only Staff readings within 6 minutes of a high water or low water were differenced.

Figure 6: (a) The root-mean-square error (RMSE) between the tide staff and the automatic gauge at Astoria Tongue-Point from 1925-1957 and (b) an estimate of the 95% confidence interval (see text for discussion). In both (a) and (b), the median monthly value for each year is given in red. The typical range of monthly values is shown by the gray shading, and is an approximation of the 25% and 75% percentile. The slightly larger values post 1945 in (b) reflect a reduction in n from ~30 per month to ~20, but also gauge issues from 1947-1949 and from Dec. 1951-Apr 1952. Additional comparison sheets are available from the EV2 database until at least 1984, but were not digitized.
Figure 7: (a) Annually averaged tide range (TR) at Tongue Point in Astoria, 1855-present; (b) seasonal variation in 30d moving average tidal range for 19th century and recent data, averaged over a nodal cycle; (c) Difference between the modern and historic tidal range ($\Delta$TR) from (b). Data from downtown Astoria (cyan and cyan crosses) has been adjusted upwards by 0.05 m to match Tongue Point data, based on nearly 4 years of simultaneous measurements from October 2015- July 2019 (see + symbols in (a) for comparison of adjusted data).

Figure 8: Change in major tidal constituent amplitudes from 1855-2019 in Astoria, Oregon, adjusted for nodal cycle variability. The 5km difference in location between the historic gage (city of Astoria, cyan coloring) and the modern gage (Tongue Point, blue coloring) makes a slight difference in some constituent values; for example, the $M_2$ and $K_1$ amplitude are 0.02 m and 0.005 m lower at Astoria than Tongue Point, respectively. Other constituent differences are 2 mm or less. A nodal correction was applied, following Pawlowicz et al. (2002). No bias corrections have been applied.

Figure 9: (a) $M_2$ admittance measured at Astoria (blue) vs. the 30d averaged water level at Portland (red); (b) rating curve of water level in Portland (x-axis) vs river flow at the Beaver Army Terminal; (c) Curve fit of measured Portland Water Level (y-axis) vs. observed $M_2$ admittance anomaly (x-axis) (d) curve fit of Portland Water Level vs. measured $M_2/M_2^2$ anomaly.

Figure 10: Validation of a model of river stage in Portland (PDX) and Columbia river discharge at Beaver based on Astoria tidal properties and the tidal discharge estimation (TDE) method of Moftakhari et al. (2013). (a) Tidally based estimates (red curve) of water levels well approximate the measured water levels in the Portland metropolitan area, including at Vancouver (black curve) and Portland (blue curve). (b) Estimate of river discharge, using results from (a) and Equation (2). Gray shading indicates the 95% confidence interval. The effective time resolution of each individual measurement is ~30 days.
Figure 11: (a) Annual hydrograph estimated using the Tidal Discharge Estimation method (1858-1876) and modern river discharge measured at the Beaver Army Terminal (Rkm 86); (b) Estimation of the water level change at Astoria caused by the estimated hydrograph, using the results from (a) and Equation (1); (c) Estimated seasonal difference in water level caused by changing river (difference between the historic and modern curve in (b)).

Figure 12: (a) Reconstructed annual mean sea level variations at Astoria, OR between 1853 and 1876, based on hourly data from 1855-1876 and high/low data from 1853-1854, adjusted for the offset between mean sea level and mean tide level (see supplementary data). Datum is the 1872 staff zero. Flow adjusted data given in cyan, and the cone of possible scenarios for the staff zero is given with grey shading (see Table S.2.3 and section S.2.4 in supplement for detailed description of scenarios). San Francisco sea-level data from NOAA is also plotted. An estimate for 1883 sea level in Astoria, based on 7 months of monthly tabulations, is given with error bars; this record is anomalously low, so is subsequently neglected because quality assurance of the record is not possible (b) Comparison of annual MSL at Tongue Point (blue) and Astoria Youngs Bay (green) between 1931 and 1942, relative to the NAVD-88 datum. The dashed blue line shows sea-level at Tongue Point before our datum correction (see section 2.3.4 and Supplement section S.2.5). San Francisco sea level shown for comparison (c) Monthly sea level at Tongue Point (blue), Hammond (red) and Astoria downtown (cyan) from 2011-2019. In (b) and (c), the average vertical offset $\Delta$ between Astoria-Tongue Point and measurements at Youngs Bay, Hammond/Ft Stevens, and downtown Astoria are given by the green, red, and cyan text. In (a) and (b), San Francisco data are expressed on an arbitrary datum.

Figure 13. Seasonal cycle of mean water level at Astoria, before (a) and after (b) correcting for river discharge (see Figure 11). A 30d moving average has been applied. The x-axis labels are applied at the middle of each month. The shading in (a) denotes the 25% and 75% bounds (interquartile range) for the historic (light grey) and modern (darker grey) periods. Note the overlap of the bounds is the darkest gray.
Figure 14. (a) Reconstructed relative sea level (RSL) at Astoria and Tongue Point, relative to the NAVD-88 datum. Measurements have been corrected for river flow and the unstable 1925-1960 datum. A mean height offset of 1.484 ± 0.02m is used to convert 19th century sea level (Fig. 12a) from staff zero to NAVD 88; (b) RSL for the Fort Stevens/Hammond combined station, relative to the NAVD 88 datum; (c) Geocentric MSL rise at the Astoria and Hammond/Fort Stevens locations, after adjusting for estimated vertical land motion rates (see text and Table 3; the adjustment is made relative to sea level from the year 2018). The river flow adjustment added to Astoria and Tongue Point data was made such that the mean adjustment from 1855-present is zero.
Table 1

Table 1. Summary of water level data and data products found, used, and/or digitized in this study. Light Grey Shading: recovered from archives and digitized. Dark Grey shading: Data product measured or calculated by this study. See Figure 1 for locations, and supplementary material for examples of archival documents. Notes below describe the source of archival material. Abbreviations: MSL = Mean Sea Level; WL = water level. See acknowledgement section for online link to data.

<table>
<thead>
<tr>
<th>Station (Station)</th>
<th>High/Low</th>
<th>Hourly</th>
<th>Tide Staff</th>
<th>Marigram</th>
<th>Miscellaneous Water Level Measurements</th>
<th>Meta Data</th>
</tr>
</thead>
<tbody>
<tr>
<td>Astoria (Rkm 24)</td>
<td>1853-1855, 1858 1860-1876</td>
<td>1870-1876</td>
<td>1854-1867 1871-1876</td>
<td>1853-1876b 1853-1876 and 1883/1884 summary sheet</td>
<td>Letters e,g,h, Notes a,e, Leveling Surveys c,e,i, Time Comparisons a,b, Water Temperature and Weather data a,b,c</td>
<td></td>
</tr>
<tr>
<td>Astoria River Pilots Dock (Rkm 24)</td>
<td>10/2015 - 07/2019</td>
<td></td>
<td></td>
<td></td>
<td>Campbell Scientific Radar Gauge, CS476</td>
<td></td>
</tr>
<tr>
<td>Astoria Tongue Point (Rkm 29.8)</td>
<td>1980-present</td>
<td>1925-present</td>
<td>1925-1957</td>
<td>--</td>
<td>Summary Sheets c,d</td>
<td></td>
</tr>
<tr>
<td>Astoria Youngs Bay</td>
<td>--</td>
<td>1931-1943a,c</td>
<td>--</td>
<td>--</td>
<td>Summary Sheets c,d</td>
<td></td>
</tr>
<tr>
<td>Fort Stevens (Rkm 15)</td>
<td>--</td>
<td>1940-1942</td>
<td>--</td>
<td>--</td>
<td>Summary sheet f including MSL for 1905-06; 1913;1926; 1936;1958; 1981</td>
<td></td>
</tr>
<tr>
<td>Hammond (Rkm 16)</td>
<td>--</td>
<td>1982-1988</td>
<td>2011-14</td>
<td>--</td>
<td>--</td>
<td></td>
</tr>
<tr>
<td>Portland (Willamette River)</td>
<td>--</td>
<td>--</td>
<td>--</td>
<td>Daily WL, 1876-1972</td>
<td>Leveling surveys</td>
<td></td>
</tr>
<tr>
<td>Vancouver (WA) (Rkm 160)</td>
<td></td>
<td></td>
<td></td>
<td>WL during spring freshets, 1872-1877</td>
<td></td>
<td></td>
</tr>
<tr>
<td>The Dalles (Rkm 307)</td>
<td>--</td>
<td>--</td>
<td>--</td>
<td>River Discharge, 1878-present. (USGS)</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Beaver Army Terminal (Rkm 86)</td>
<td></td>
<td></td>
<td></td>
<td>River Discharge (daily, with 30d averaged resolution)</td>
<td>See section 2.4 for details.</td>
<td></td>
</tr>
</tbody>
</table>

Notes: The archival records above were obtained from
(a) NOAA headquarters in Silver Spring, Maryland; these records are typically stored at the Federal Records Center in Suitland, Maryland.
(b) United States National Archives in Kansas City, Missouri; these archives are stored at the Federal Records Center in Lee’s Summit, Missouri. Marigrams for the year 1854 are missing, as are the rolls for Nov. 1858, May 1862, and May 1868. Due to time limitation, not all marigrams in 1870s were photographed.
(c) The EV2 database at the National Centers for Environmental Information (NCEI) (https://www.ncdc.noaa.gov/EdadsV2)
(d) NOAA Center for Operational Oceanic Products and Services (NOAA CO-OPS)
(e) United States National Archives II, College Park (Maryland)
(f) City of Portland (archives or Bureau for Environmental Services); also, Signal Service Archives at NCEI
(g) United States National Archives in San Bruno, California
(h) United States National Archives in Seattle, Washington
(i) United States Geological Survey Archives, Colorado
(j) Burgette et al. (2009)
(k) Wilson, J.M. (1878)
(l) Archival tabulations including river stage (1949-1972) and water temperature (1949-1961) from personal communication, Jason Cooper, NCEI, October 16, 2015
(m) The National Weather Service district library, Portland, Oregon
Table 2

Table 2: Estimated uncertainty bounds for Astoria annual sea-level data from 1853-1876, relative to NAVD-88 datum. Uncertainty bounds includes estimates of historical leveling error and the cone of uncertainty in datum caused by different interpretations of leveling data, the uncertainty in datum tie to the NAVD-88 datum, and the precision of the measurement. The various sources of uncertainty are combined under the assumption they are uncorrelated. See supplement Section S.2 for more information on evaluation of uncertainty (and section S.2.7 for a synopsis).

<table>
<thead>
<tr>
<th>Uncertainty (± m)</th>
<th>1853-1855</th>
<th>1856-1858</th>
<th>1859-1865</th>
<th>1866</th>
<th>1867-1876</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>(0.07, -0.07)</td>
<td>(0.03, -0.07)</td>
<td>(0.06, -0.03)</td>
<td>(0.04, -0.03)</td>
<td>(0.03, -0.03)</td>
</tr>
</tbody>
</table>

Notes:

(a) Leveling surveys post 1858 estimated to be accurate to within ± 0.01 m, based on level loops in 1876 and 1887.
(b) Larger uncertainty pre-1858 reflects documented subsidence in staff and gauge, and inferred lower accuracy in leveling (see text).
(c) Leveling surveys in 1861 and 1862 disagree by 0.05 m, producing greater uncertainty bounds from 1858-1866
(d) An uncertainty of ± 0.02 m is estimated for the datum tie between the 19th century record and the NAVD-88 datum.
Table 3: Estimates of sea-level rise since 1858, with and without a correction for vertical land motion (VLM). The linear trend from 1926-2014 is also reported, based on the period of overlapping data for the three sea level stations listed below.

<table>
<thead>
<tr>
<th></th>
<th>Relative Sea-Level Rise (1858-1876 to 2000-2018 epoch)</th>
<th>Trend in Relative Sea Level, 1926-2014 (mm/yr)(a)</th>
<th>Vertical Land Motion (mm/yr)</th>
<th>Geocentric Sea Level Rise from 1858-1876 to 2000-2018 epoch after correction for vertical land motion</th>
</tr>
</thead>
<tbody>
<tr>
<td>Astoria</td>
<td>0.06 ± 0.04m</td>
<td>0.25 ±0.14 (N=89)(b)</td>
<td>0.35 ± 0.15(^{c,h})</td>
<td>0.384 ± 0.6(^{d,h})</td>
</tr>
<tr>
<td>Fort Stevens</td>
<td>--</td>
<td>-1.0 ±0.34 (N=16)</td>
<td>1.29 ± 0.37(^c)</td>
<td>1.56 ± 0.47(^d)</td>
</tr>
<tr>
<td>San Francisco</td>
<td>0.21m</td>
<td>1.88 ±0.15 (N=89)</td>
<td>-1.3 ± 0.6(^{c,f})</td>
<td>0.024 ± 0.09m (0.061 ± 0.09m)</td>
</tr>
</tbody>
</table>

Notes

a. Trends based on time period of available records at the combined Fort Stevens/Hammond station.

b. The calculation is based on discharge corrected, datum corrected data. The rate obtained for uncorrected data is -0.31 ± 0.19 mm/yr.

c. Based on estimates of VLM from sonel.org, a repository of GNSS records near tide gauges that was developed under the auspices of the Global Sea Level Observing System (sonel.org; see e.g., Gravelle et al., 2013),

d. Based on estimates of VLM from the Nevada Geodetic Library using the IGS-08 reference datum; methods described in Blewitt et al., 2016 and 2018. Rates were downloaded from [http://geodesy.unr.edu/PlugNPlayPortal.php](http://geodesy.unr.edu/PlugNPlayPortal.php). Rates were updated for measurements through August 2019.

e. Estimate based on GNSS station FTS6; other values for Fort Stevens from GNSS sensor FTS5. Records from 1996 to 2019 used.

f. Based on GNSS station TIBB, the standard VLM measurement used for San Francisco (e.g., NRC, 2012). The sensor is located nearly 10km from the San Francisco gauge. VLM estimate based off 1994-Aug. 2019 data. An earlier estimate of the VLM at this station, using records through mid-2018, was -1.06 mm/yr (+0.24 mm/yr larger than listed in table above). This indicates some uncertainty in the VLM at this station.

g. Geocentric sea-level rise after including Zervas (2009) datum correction.

h. GNSS station TPW2, from data available from 2000 to 2019.
Figure 2.
Figure 3.
Figure 4.
(a) Hourly - Staff Reading
1855-1859

(b) Hourly - Staff Reading
1860-1867

(c) Hourly - Staff Reading
1871-1876
Figure 5.
(a) HW - Staff Reading
1860-1876

(b) LW - Staff Reading
1860-1876

\[ \mu = 0.015m \]
\[ \sigma = 0.015m \]

\[ \mu = 0.006m \]
\[ \sigma = 0.013m \]
Figure 6.
(a) RMSE between staff and gauge

(b) Confidence in Monthly Mean
Figure 7.
(a) Tidal Range

(b) MTR = 2.08 m

(c) Δ Tidal Range:
Modern - Historical
Figure 10.
Figure 11.
(a) Estimated Discharge

Average $Q = 6.5 \times 10^3 \text{ m}^3/\text{s}$

Average $Q = 7.8 \times 10^3 \text{ m}^3/\text{s}$

(b) Seasonal Water Level due to Discharge

(c) $\Delta$ Water Level
Figure 13.
RSL (m)

(a) RSL 1858-1876: 1.39 +/- 0.04 m
RSL 2000-2018: 1.45 m

(b) Fort Stevens

(b) Hammond

(c) VLM corrected

1860 1880 1900 1920 1940 1960 1980 2000 2020

1.3 1.4 1.5

RSL (m)

1.3 1.4 1.5

MSL (m)