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# Po River discharges: a preliminary analysis of a 200-year time series

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**Abstract** Daily stage data measured at the closing section of the Po River (Northern Italy), collected from historical archives of the "Hydrological Office of the Po River—Parma", allowed to estimate daily discharges for the period 1807–1916, therefore to extend the time series of Po River discharges for 110 years before the actually published time series. This paper provides a detailed description of the stage–discharge conversion and of the tests performed for checking homogeneity of reconstructed data. In particular, monthly discharge data were compared with approximated catchment-average data of precipitation and evapotranspiration for the period 1831–2003, which were estimated, respectively, from monthly data of local precipitation (at Milan, Turin and Parma) and local air temperature (at Milan and Turin). It emerges that estimated values of precipitation, evapotranspiration and discharge provide a coherent picture of the hydrological dynamics in the basin throughout the study period. Specifically, an apparent progressive depletion of basin reservoirs is observed since 1920, i.e., when concomitant sudden changes (regime shifts) occurred in precipitation (downward shift) and evapotranspiration (upward shift). The 1920-shift is amongst the likely causes of the fact that prolonged drought periods as those observed in the 1940s and since 2003 are not observed in the pre-shift period, when accumulation of reservoirs occurred. The increase in peak-flow discharges observed in recent decades, with values well above the maximum discharge estimated for the nineteenth century as well as for the early twentieth century, is apparently the result of the massive levee works along the river network that were completed in the 1960s. On decadal time scales, discharge variability is found to essentially reflect the changes in precipitation patterns. In particular, peaks of comparable magnitude are found in the 128-month (~11 years) wavelet spectra of precipitation and discharge. Furthermore, concurrent changes are observed in the persistence (i.e., autocorrelation) of precipitation and discharge data. Since the red-noise background spectrum of discharge is much lower than that of precipitation, river discharge

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is the likely hydrological variable to be preferred for assessing the basin's response to the background climatic variability occurring at the decadal and multi-decadal time scales, notwithstanding the fact that changes in water management and other anthropogenic impacts can be important on long time scales. Concerning the seasonal to interannual response to climatic forcing, a robust dependence of wintertime precipitation and discharges on the state of the NAO was observed. This dependence results in stronger (weaker) precipitation and higher (lower) discharges during negative (positive) anomalies of the NAO index.

## 1 Introduction

### 1.1 Motivation

Discharge is essentially a function of the spatial integration of precipitation over a river catchment. The discharge property of smoothing out precipitation variability has contributed to an increasing use of discharge records in the field of climatology, especially of data concerning rivers with large-scale basins (e.g., greater than $10^4$ km$^2$) for the identification of forcing mechanisms at the hemispheric to global scales. The availability of long discharge time series and the increasing number of nations publishing discharge related information have further contributed to focus on the use of discharge records in climatological investigations, which led to the constitution of global networks for the collection of discharge estimates (e.g., Fekete et al. 2002).

Actually, discharge data reflect also all the dynamic structures of the physical processes linking precipitation to discharge, these processes depending on natural non-linear controls largely influenced by climatic factors (such as evapotranspiration, infiltration, snowpack dynamics, etc.) and on various human interventions (such as channel and catchment changes, reservoirs management, etc.). Therefore, the interpretation of river flow variability cannot neglect the understanding of the land-surface hydrologic system, which is generally gained by complementing quantitative observations with outputs of climate and hydrological models (e.g., Milly and Wetherald 2002).

However sophisticated the model approach may be, the availability of hydrological data is a necessary prerequisite to characterize the hydrological variability in a river basin and to uncontroversially address its dependence on climatic fluctuations. Furthermore, while high-frequency river-flow variability can be interpreted only by using a detailed modelling of the surface drainage network, the dominant physical controls of hydrological variability at decadal or longer timescales can be identified by a comprehensive analysis of continuous, long-term geophysical data (e.g., secular time series of temperature, precipitation and discharge).

This work was inspired by the increasing relevance gained by hydrological data—in particular discharge data—for assessing and interpreting the repercussions at the regional/local scale of global climate conditions. In fact, although nowadays global climate model scenarios clearly point to sharpening of extremes as well as to changes of seasonality in regional temperature and precipitation (Christensen et al. 2007), some controversy still exists over whether, and eventually how, global change controls local hydrological regimes and water resources (e.g., Helmiö et al. 2005). As a consequence, a series of questions needs to be addressed: is the impact of climatic change already detectable in river discharge data? Does river discharges show trends or other significant changes of important characteristics? Are these changes compatible with natural variability or are they related to anthropogenic influence?

In this connection, this work is motivated by the concern originated from the hydrological extremes of exceptional severity—both floods and droughts—that have been observed since the last decade in the Po River basin (Northern Italy). Specifically, it is motivated by the need to assess the decadal-to-centennial climatic and hydrological variability, which is the background upon which these events occurred, and to address how much the present-day conditions are the result of natural variability an how much they are the result of anthropogenic influence in the area. In this connection, given the availability of long time series of local rainfalls and air temperature for Northern Italy (e.g., Cati 1981; Camuffo 2004; Brunetti et al. 2006), the evaluation of a long time series of Po River discharge could help the understanding of hydrological response of the catchment to climate dynamics.

### 1.2 Objectives

The main objective of this study is to provide a basis for the long-term characterization of Po River discharge variability, which was gained over a 200-year period by re-evaluating daily discharge data from historical daily water-surface elevation (stage) data. This objective was achieved thanks to the preservation and availability of historical records and documents concerning the Po River, which are amongst the most valuable outcomes of the great effort devoted to understand and control the hydrological dynamics in the area (see the 'Sources of historical information' in the 'References' section). In fact, the Po River basin is the natural background which strongly contributed to the socio-economic development of Northern Italy. When the local and regional authorities in charge of planning and managing the water resources of the Po basin were found inadequate for the preservation of the environment and a sustainable development of the area, the technical power was centralized into a unique basin authority: the civil *Magistrato* for general works in the Po system. Since the set up of the *Magistrato* in 1806, punctual daily stages have been measured from the hydrometer located in the river's closing section at Pontelagoscuro (see Fig. 1).[1]

So far, the 1917–present stage data have been converted into daily discharge estimates on a regular basis, while only sporadic discharge estimates are available for the preceding period. Here, a punctual stage–discharge conversion was performed on the 1807–1916 data. Devoting emphasis on the monthly timescale, discharge data are compared with estimates of catchment-averages of precipitation and evapotranspiration in order to ascertain that the time series display a reasonably coherent picture of the hydrological and climatic variability of the nineteenth and early twentieth centuries.

## 2 The Po River: an overview

The drainage basin of the Po River (Fig. 1) extends over an area of about 75,000 km$^2$, about one-third of which is the alluvial plain of the river known as the Po Valley. The Po Valley was progressively formed by deposition of sediments that were drained from mountain chains mainly originated from the Tertiary Era corrugations: the Alps, that extend in an arc along the northern border and reach elevations well over 4,000 m, and the Apennines, that rise along the south-western border. After the maximum extent attained by the floodplain of the river in the Quaternary Era, the present shape of the Po Valley was defined by slow

[1] Pontelagoscuro is located 90 km upstream the sea and is the last non-tidal section.

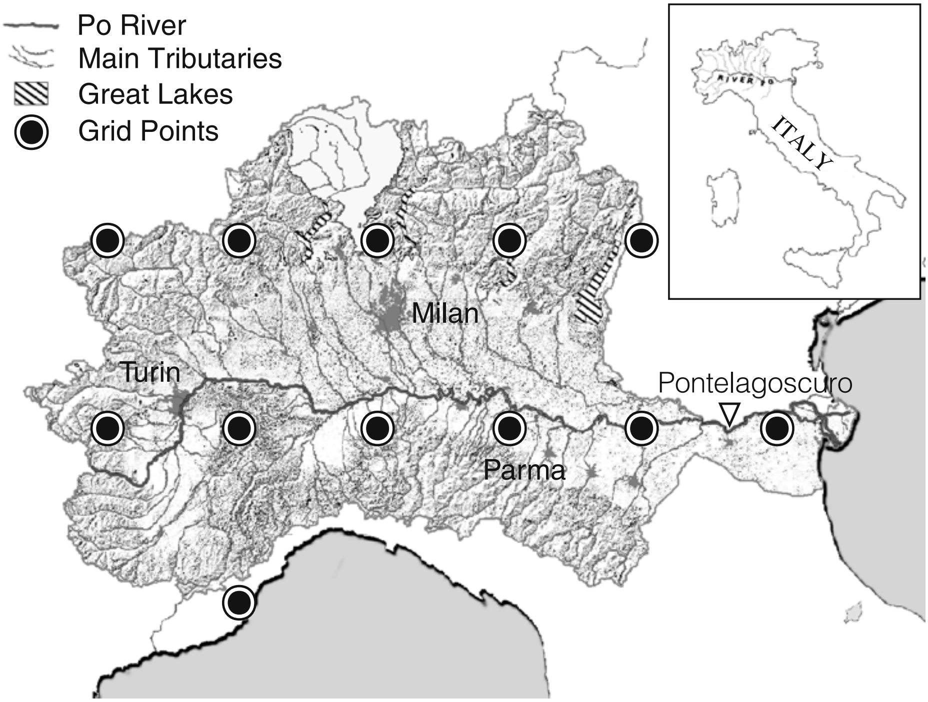


**Fig. 1** Map of the Po River catchment. The map provides the location of the Pontelagoscuro gauge, of Turin, Milan and Parma and of the grid-points of rainfalls data by Brunetti et al. (2006)

processes of subsidence and by the progressive rise of the sea level (e.g., Zanchettin et al. 2007), that resulted in a retreat of the shoreline in the proximity of the river mouths as well as in a broader extension of marshes and lagoons along the littoral zone of the northern Adriatic Sea.

The present-day network of the Po River is commonly divided into four main stretches (e.g., Marchi et al. 1995): the Upper Po (with a drainage area of approximately 37,000 km$^2$), the Middle Po (68,000 km$^2$), the Lower Po (70,000 km$^2$) and the Po Delta. The river morphology has evolved from a geometry with irregular meandering channels controlled by discontinuous levee alignments and riverbank protection (early nineteenth century) to a geometry with artificial meandering or straight channels that are controlled by flood corridors and continuous riverbank protection (e.g., Govi and Maraga 2005). The present levee system, which was completed during the 1960s by the "*Magistrato per il Po*",[2] is a flood canal along the final 420 km of the Po River watercourse (including also the final stretches of its tributaries), with a remarkable channel storage capacity in the Middle Po stretch. The major outcomes of the progressive expansion of existing levee systems are a decreasing flood proneness of the Upper Po sub-basins (e.g., Sesia, Tanaro and Ticino rivers) and an increasing vulnerability of Po Plain areas (Lower Po, Po Delta) to hydrological hazards, which is seen as an increasing value of the flood peak with given probability (Marchi et al. 1995).

[2] The "*Magistrato per il Po*" is the peripheral branch of the Ministry of Public Works set up in 1956 in charge of all the technical activities concerning river training, regulation and protection of nearby lowlands.

The annual regime of the Po River, which is characterized by two low-water periods (winter and summer) and two flood periods (late fall and spring), is strongly influenced by the seasonal pattern of precipitation. The first flood period reflects the intensification of rainstorms in late fall, while the second flood period reflects the contribution of snowmelt processes in the most elevated portions of the catchment (Cattaneo et al. 2003).

Precipitation displays a wide variety of patterns resulting in a complex spatial–temporal heterogeneity (Quadrelli et al. 2001), which is partly originated by the orography of the region strongly affecting atmospheric dynamics on a broad range of scales, and particularly on mesoscales (Bertò et al. 2004). Three main sub-regions are identified in a classical subdivision of the Po River basin area in function of the spatial distribution of rainfalls (e.g., Cati 1981): Piemonte (36,356 km$^2$), Lombardia (21,074 km$^2$) and Emilia (12,661 km$^2$). These sub-regions contribute differently to the total discharge of the Po, especially concerning high discharges (see Table 1).

Notably, at the monthly (or preferably seasonal) time scale, precipitation over Northern Italy shows a well-identifiable response to jet-streams and to the large-scale atmospheric patterns (teleconnections) upon which jet-streams depend. Specifically, precipitation variability is likely modulated by the North Atlantic Oscillation—or NAO—(e.g., Quadrelli et al. 2001; Tomozeiu et al. 2002), which is a meridional oscillation in atmospheric masses between the Arctic and the subtropical Atlantic affecting the transport and convergence of atmospheric moisture over the North Atlantic area (e.g., Hurrell 1995; Marshall et al. 2001).

Although discharge fluctuations are dominated by precipitation, land-surface processes (evaporation/evapotranspiration, infiltration) and human management of groundwater reservoirs (mainly withdrawn for drinking use) alter the low-frequency response of discharges to precipitation. Specifically, only 60% of the total annual inflow, which has been estimated in about 77.7 × 10$^9$ m$^3$ (i.e., the annual mean precipitation is about 2,465 m$^3$/s), is converted, on average, into discharges (Cati 1981; IS 2005). Evapotranspiration is the major process intervening in the conversion of precipitation signals into discharges; other detectable components of evaporation processes, such as water-surface evaporation from lakes, contribute much less to the overall budget in the basin. For instance, the estimated annual-average total loss of water mass from the Lake Garda (the largest Italian lake) caused by evaporation is about 1/1,000 of the annual-average total outflow of the Po River (Cati 1981).

Furthermore, natural hydrological features and human management of large-scale water bodies (i.e., the deep sub-alpine natural lakes) can strongly influence discharge variability. Actually, sub-alpine lakes (Como, Garda, Iseo, Idro and Maggiore, for a total water mass volume of about 1.25 × 10$^9$ m$^3$) are regulated according to reservoir management strategies and provide an average total outflow of about 573 m$^3$/s (Premazzi et al. 2003).

**Table 1** Contributions of precipitation over three sub-regions of the Po River basin to the maximum discharge of the Po at Pontelagoscuro during four floods in the twentieth century (after Cati 1981)

| Region | Total precipitation ($1E + 6$ m$^3$ and %) | | | | | | | |
|---|---|---|---|---|---|---|---|---|
| | 1917 | | 1926 | | 1928 | | 1961 | |
| Piemonte | 7,712 | (62) | 6,499 | (57) | 9,157 | (51) | 9,620 | (58) |
| Lombardia | 4,215 | (34) | 4,063 | (35) | 6,767 | (37) | 4,847 | (29) |
| Emilia | 506 | (4) | 921 | (8) | 2,204 | (12) | 2,076 | (13) |
| Total | 12,433 | | 11,483 | | 18,146 | | 16,453 | |

Recent assessments (IS 2005) have quantified the total annual derivations for industrial use in about $7.8 \times 10^9$ m$^3$ and for agricultural use in $21.9 \times 10^9$ m$^3$ (i.e., an average value of 694 m$^3$/s). Agriculture is the main land use over the Po Plain, which explains the huge total length of the network of artificial canalisations over the basin (about 16.750 km), and in particular the impressive 85 km-long Cavour irrigation canal, which was opened in 1866 and diverts up to 110 m$^3$/s from the Upper Po to the Ticino.

Considering that the average Po River discharge is about 1,500 m$^3$/s (Cati 1981), the ensemble of these contributions can considerably alter the response of discharges to precipitation.

### 2.1 The Po at Pontelagoscuro: historical background

The archives of the "Hydrological Office of the Po River—Parma" (HOPR) constitute an essential source of information for a preliminary investigation on historical characteristics of the Po River. Information is mainly available in the form of internal reports and national publications (i.e., 'Annual Acta of Public Works'). In particular, the publication N.1539 (HOPR 1935) collects the daily stage data measured in the period 1807–1930, with a detailed description of the implemented procedures for data homogenisation. Several Authors (e.g., Fantoli 1912; Giandotti 1927; Visentini 1938) studied these hydrometric observations and the factors that may had affected the stage–discharge relationship at Pontelagoscuro since the early nineteenth century. Amongst others, the undated, internal report of the HOPR by Giovannelli and Allodi (196X) summarizes the knowledge achieved by the HOPR about the Po at Pontelagoscuro for the period 1917–1941,[3] which is a period of considerable interest as it matches the time when the HOPR started to regularly convert stage measurements into discharge estimates by using a discharge rating curve (DRC).

Historically, the earliest known DRC of the Po at Pontelagoscuro was defined by *Lombardini* according to the set of discharge estimates reported in Table 2; this DRC was later modified by *Possenti* until the final form dated 1867 (MPW 1878). In 1878–1880 the *Brioschi Commission* defined the first empirical DRC established by using modern gauging techniques and a complete set of hydrometric observations (Betocchi 1881). The 'Brioschi Curve' was dropped when a new set of hydrometric measures (conducted by the HOPR between 1923 and 1925) showed an overestimation of discharge values (Giovannelli and Allodi 196X). The new DRC was therefore used by the HOPR to re-estimate daily and monthly discharges for the period 1917–1922 (i.e., since the previous flood which occurred in 1917).

A further backward application of the HOPR's DRC must be done with caution: such stage–discharge conversion may lead to considerable uncertainty in the discharge estimation even when historic stage measurements are accurate and homogeneous. In fact, changes in physical features (such as the water-surface slope and the cross-section shape) and possibly other factors (such as land subsidence) could have caused substantial modifications to the stage–discharge relationship, hence to the DRC. In particular, floods and other high water events (such as those of 1839, 1857, 1872 and 1917) could have significantly and permanently affected the channel geometry in the vicinity of the gage. Nonetheless, Giovannelli and Allodi (196X) formerly used the HOPR's DRC to estimate a number of nineteenth century peak-flow discharges (above 7,000 m$^3$/s) from historical

[3] The report is undated, although realistically it was written in the early 1960s (Allodi, personal communication).

**Table 2** Water-surface stage measures and discharge estimates before 1878

| Date | Water-surface stage | Discharge | | |
|---|---|---|---|---|
| | *m* respect to the hydrometric zero | Various experimentators $m^3/s$ | Possenti $m^3/s$ | HOPR $m^3/s$ |
| 19 Dec. 1811 | −3.83 | 1,110 | 1,369 | |
| 29 Nov. 1811 | −3.63 | 1,221 | – | |
| 1812 | −2.84 | 1,693 | 2,073 | |
| 1815 | 1.84 | 4,740 | 5,272 | |
| 1820 | −2.36 | 2,176 | 2,176 | |
| 1839 | 2.96 | 7,193 | – | 7,120 |
| 1846 | 2.6 | 6,382 | | – |
| 1857 | 2.96 | 7,532 | | 7,120 |
| 1866 | 2.98 | | | 7,170 |
| 1868 | 3.05 | | | 7,320 |
| 1872 | 3.32 | | | 7,930 |

The last two columns report the discharge values estimated by applying the discharge rating curves by Possenti and by the HOPR. All data are from Giovannelli and Allodi 196X.

stage records (see Table 2). Their work suggests that, with the due caution, the backward application of the HOPR's DRC is a viable proposition.

A number of facts concerning the Pontelagoscuro section supports the assertion that it is a reasonable assumption that the stage–discharge relationship has not changed significantly on long-term since the oldest hydrometric measures. In particular, scour and deposition processes during extreme flooding episodes apparently had only a temporary effect on the channel geometry. For instance, after the excavation of about 8 m caused by the flood of May 1905 (CNI 1905) the cross-section was reported to revert to the original shape within a few weeks (Canali 1959). Furthermore, only minor long-term variations were observed for the area and the average water-surface slope below the stage for an ordinary drought, as well as for the average water-surface slope at flooding level (see Table 3). Finally, the river-bed elevation above the mean sea level is reported to have been comparatively stable throughout the nineteenth century (Cati 1981) although the Po Plain is, generally, affected by land subsidence (Carminati and Di Donato 1999).

**Table 3** Historical information about factors which may had altered the stage–discharge relationship of the Po at Pontelagoscuro since the early nineteenth century: area ($m^2$) (column A) and average water surface slope (column B) below the stage for an ordinary drought ($H = -4.53$ m); average water surface slope at flooding level (column C)

| Year | A | B | C |
|---|---|---|---|
| 1812 | – | – | 0.11‰ |
| 1872 | – | – | 0.10‰ |
| 1874–1875 | 1,208 | 0.07‰ | – |
| 1905 | 1,260 | 0.07‰ | – |
| 1917 | – | – | 0.12‰ |
| 1923–1925 | 1,190 | 0.07‰ | – |
| 1926 | – | – | 0.11‰ |

Source: Giovannelli and Allodi 196X

## 3 Data

This study makes use of the following dataset:

- the daily stages of the Po River at Pontelagoscuro (hereafter: stages) between 1807 and 1930. The original data were digitalized from the archival publication N.1539 of the HOPR.
- the daily Po River discharges at Pontelagoscuro (hereafter: discharges) since 1917. The data were provided by the Regional Agency for Environmental Protection (ARPA)—Emilia Romagna, Hydro-meteorological Office. Discharge estimates above 1,000 $m^3/s$ are rounded to the nearest tenth;
- the 1865–2003 monthly sequences of rainfall grid-data provided by the ISAC-CNR Historical Climatology Group (Brunetti et al. 2006); data are rainfall anomalies (relative deviations) with respect to the 1961–1990 mean. Figure 1 shows the location of the twelve 1 × 1° grid-nodes considered;
- the updated monthly time series of local precipitation at Milan (hereafter: Mi; since 1800), Parma (Pr; since 1831) and Turin (To; since 1800), originally produced by the HOPR (e.g., Cati 1981).
- the monthly time series of land surface temperature (LST) at Turin (data from: ARPA—Piemonte) and Milan-Brera (Brunetti et al. 2006) since 1800. Data are pre-processed for the removal of the heat-island effect, which is modelled as quadratic polynomial fit to LST data after 1950 and results in an estimated increase of about 1.9°C at Milan and of about 1.8°C at Turin between 1950 and 2003;
- the monthly NAO index, which is the updated monthly time series of the Gibraltar-Iceland NAO index, originally produced by Jones et al. (1997).

## 4 Results and discussion

### 4.1 Definition of reference time series

The water balance equation in a river basin for a given period can be simplified as:

$$P = E \pm \Delta R + D \tag{1}$$

where $P$ is precipitation, $E$ is evapotranspiration, $R$ is water reservoirs (soil and groundwater, snowpack, lakes and so forth) and $D$ is runoff which eventually produces discharge. An accurate assessment of the hydrological balance would lead to an ideal reference time series of the form: $D_{\mathrm{ref}} = P - E \pm \Delta R$, which would be suitable for testing the long-term consistency of independently estimated discharge data. Actually, approximations are inevitably introduced when evaluating historical catchment-averages of $P$, $E$ and $R$, which must be accounted for when assessing data homogeneity. Here, lacking necessary data it is assumed that the contribution to runoff of long-term changes in water reservoirs is negligible with respect to $P$ and $E$ (i.e., $\Delta R \sim 0$). Accordingly, the long-term water balance is approximated by: $D_{\mathrm{ref}} = P - E$.

The time series of $P$ and $E$ are evaluated as follows.

($P$) Monthly regional precipitation inflows (i.e., precipitation multiplied by surface area) from Piemonte, Lombardia and Emilia are calculated from monthly local precipitation at Turin, Milan and Parma, respectively, assuming that each of these stations is representative of the average rainfall conditions over its belonging region. The viability of this approach seems supported by the good correspondence observed between local precipitation data and regional-average rainfalls obtained by grid-data (see Fig. 2).

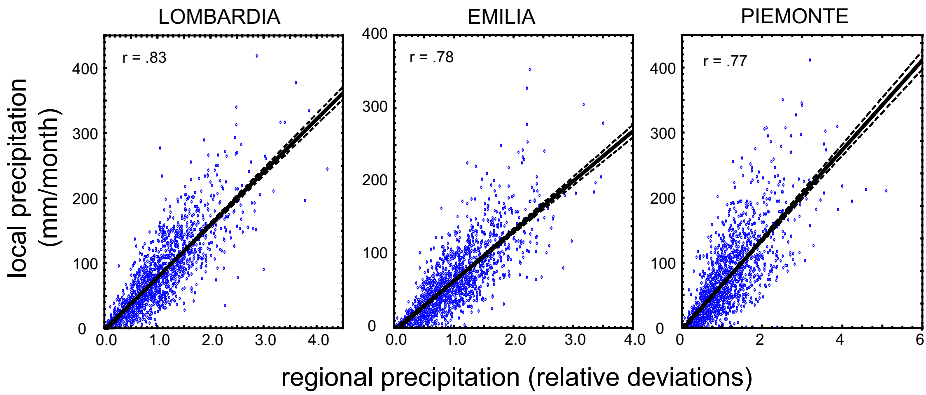


**Fig. 2** Scatterplots and best-fit curves of linear regression (with 95% confidence bands) between the regional-average precipitation estimated from grid-data (*abscissas*) and local precipitation at representative locations (*ordinates*): Milan/Lombardia, Turin/Piemonte and Parma/Emilia. Pearson's coefficients of linear correlation ($r$) are also provided

Concerning the period 1831–2003, the estimated average annual inflows in Piemonte, Lombardia and Emilia is $30.9 \times 10^9$, $20.9 \times 10^9$ and $9.9 \times 10^9$ m$^3$, respectively. The ensemble of these regional contributions ($61.7 \times 10^9$ m$^3$) underestimates the expected total annual inflow (which is $77.7 \times 10^9$ m$^3$, as detailed above) for about 21%. In order to correct this discrepancy, the monthly catchment-average precipitation estimates are corrected according to equation: $P^* = 1.26P$.

($E$) Potential evapotranspiration (PET) can be calculated from monthly LST data through the Thornthwaite formula (Thornthwaite 1948). Although actual evapotranspiration depends on complex interactions between meteorological conditions, vegetation and soils, while PET depends only on LST, this approach seems suitable for the objectives of the present study.

In order to estimate the monthly loss of water over the basin associated to PET, regional-average LST is estimated from local LST at Milan and Turin. The viability of this approach is supported by the remarkable uniformity observed in LST data over the basin (Cati 1981), which differ mainly for an altimetry-related gradient of about −1°C/190 m. So, the annual LST regime for a given altimetric zone can be calculated on the basis of this gradient by using reference values at 500 masl for the annual-average (10.8°C) and for the standard deviation (8.0°C) (Cati 1981). LST data at Milan (122 masl) and Turin (269 masl) provides results that are coherent with this indication: 1800–1949 (i.e., before heat-island effects) averages±standard deviations are, respectively, 13.0 ± 8.1°C and 11.9 ± 8.0°C.

Local LST data are decomposed according to the additive model $X_t = \mathrm{TC}_t + \mathrm{SC}_t + I_t$, where $t$ stands for the particular point in time, TC is the trend-cycle component, SC is the seasonal component (i.e., regular oscillations with given periodicity) and $I$ is the irregular component. According to the Census 1 procedure (Makridakis and Wheelwright 1989), seasonal decomposition is performed through the following steps: the original series is averaged by applying an unweighted moving-average filter with window equal to the chosen seasonality (12 months, here); the difference between the original and the averaged series isolates SC plus $I$; SC is then evaluated as the average for each point in the season (each month of the year, here); finally, the seasonally adjusted series (TC + $I$) is calculated by subtracting SC from the original series. Figure 3 illustrates the annual regime and the seasonally adjusted series extracted from the LST data at Milan and Turin.

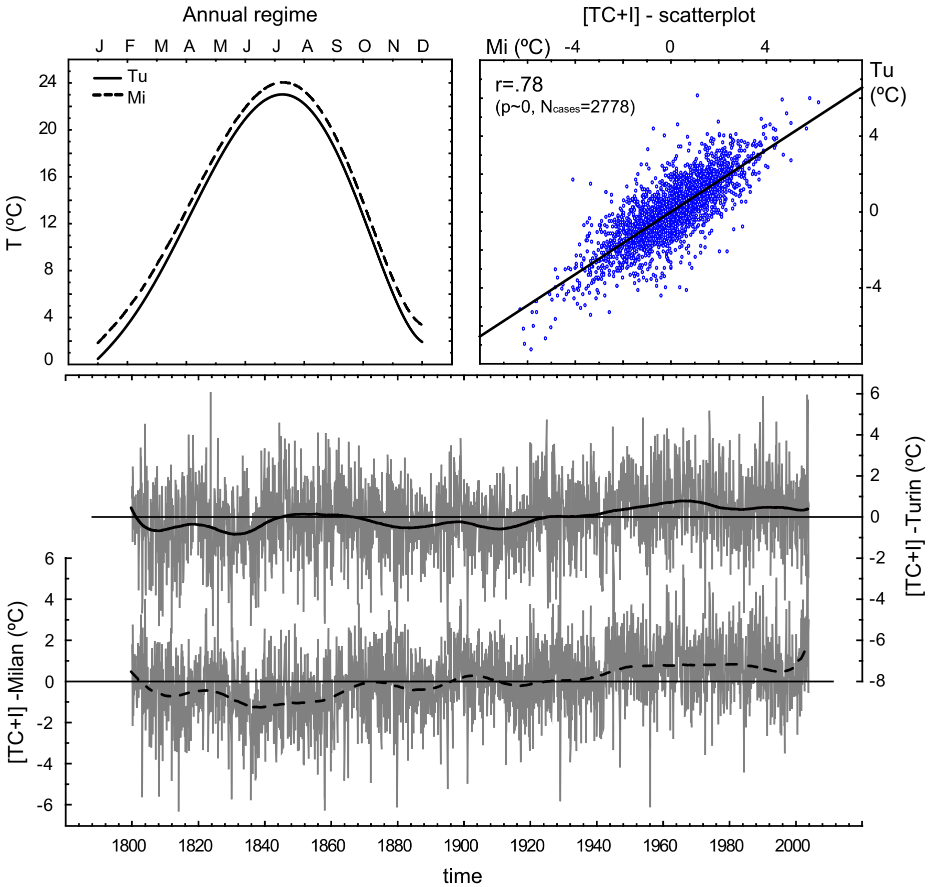


**Fig. 3** Annual regime (12-month seasonal factor) and seasonally-adjusted series (TC+*I*) extracted from land surface temperature data at Milan and Turin for the period 1800–2003. The *black lines* in the *bottom panel* are the exponentially-weighted fits to the seasonally-adjusted series, for which the weights of individual data points for consecutive segments of the curve are calculated according to a negative exponential function

LST estimates for a given altimetric zone are evaluated by adding the Milan-Turin average (TC + *I*) component to the SC component specific of that zone. Table 4 illustrates the three altimetric zones and the associated LST regimes into which the Po basin area has been divided. Here, it is assumed that the contribution of mountain areas (i.e., above 600 masl with climate typologies ranging from temperate-cold to cold and snowy) to total PET is negligible with respect to that of flat areas where agriculture is the dominant land use.

**Table 4** Characterization of altimetric zones in the Po River basin

| Range (masl) | *T* regime (°C) | Extent (% of territory and km$^2$) | | | | | |
|---|---|---|---|---|---|---|---|
| | (mean±SD) | Piemonte | | Lombardia | | Emilia | |
| 0–300 | 13 ± 8 | 26.4 | 9,598 | 47.0 | 9,905 | 47.8 | 6,052 |
| 300–600 | 11 ± 8 | 30.3 | 11,016 | 12.5 | 2,634 | 27.1 | 3,431 |
| >600 | | 43.3 | 15,742 | 40.5 | 8,535 | 25.1 | 3,178 |

The estimated annual-average total loss of water associated to PET is about $32.3 \times 10^9$, which leads to an $E/P$ ratio (0.41) consistent with the estimated volume loss in the conversion of the total annual inflow into discharges (Cati 1981; IS 2005).

Although the intrinsic limitations of the approaches adopted for the evaluation of catchment-averages of $P$ and $E$, the internal consistency and the physical plausibility of these results indicate that these estimates are valuable data for testing the long-term homogeneity of the discharge time series. The monthly time series of $P$ and PET are graphically presented in Fig. 4. Notably, $P$ closely traces the basin-average precipitation pattern gathered from grid-data, except deviations around the 1920s and the 1960s. The most prominent characteristic of the $P$ pattern is the rhythmic succession of peaks and valleys observed after 1920, which defines apparent cycles of about 20 years.

The hypothesis of a regime change around 1920 is tested by performing a non-parametric Cumulative Sums (CUSUM) tests for shift detection (Taylor 2000) on the average as well as on the 12-month running standard deviation (SD) of deseasoned $P$ data. Results indicate an apparent shift in the average above the 95% significance level occurring around 1920.9, this shift accounting for an estimated decrease of about 8% (comparable results are found for the basin-average precipitation anomalies estimated from grid-data: shift at 1920.9, the estimated change is −6%). A downward shift in $P$–SD is detected around 1907, which accounts for an estimated decrease of precipitation variability in the

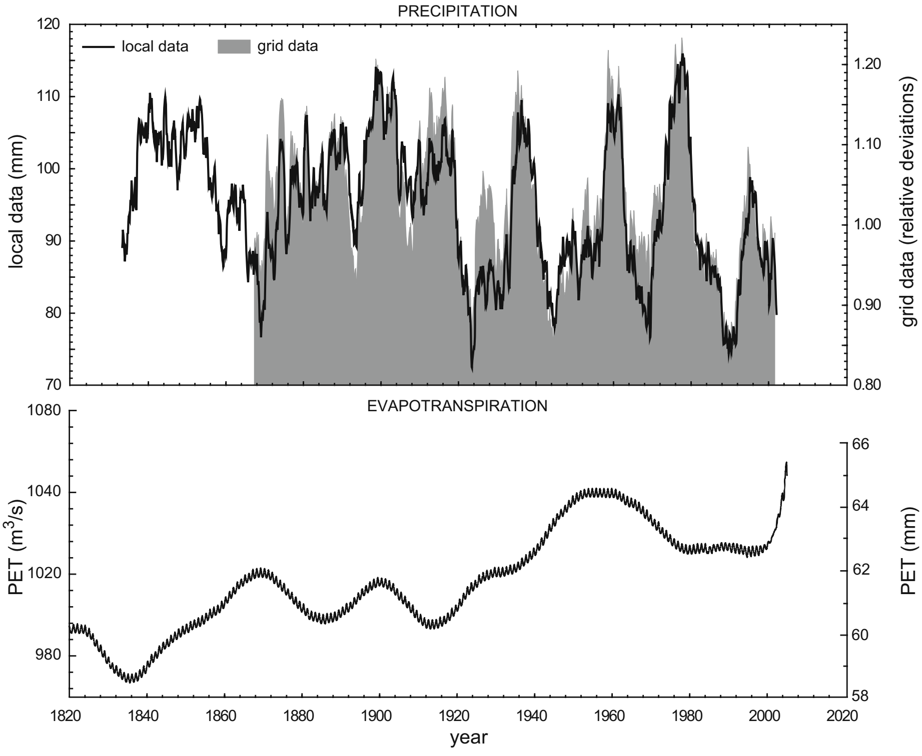


**Fig. 4** *Top panel* five-year running means of basin-average monthly precipitation over the Po River basin as estimated from local data and from grid data. *Bottom panel* distance weighted least-square fitting to the monthly time series of potential evapotranspiration (PET) over the Po River basin

post-shift period of about 21% (when the CUSUM test is based on the rank instead of on values, this shift is not statistically detectable).

Similarly, CUSUM tests for shift detection are performed on deseasoned *E* as well as on deseasoned (*P*–*E*) data. Concerning *E*, a prominent regime shift in the average is observed at 1942.4 (+4%), with secondary shifts at 1920.2 (+1%) and 1966.5 (−2%); no shifts are observed in *E*–SD. Concerning (*P*–*E*), downward shifts are detected at 1920.9 in the average (−16%) and around 1907 in the SD (−23%).

### 4.2 Estimation of daily discharges from stage data

The top panel of Fig. 5 summarizes the relationship between daily stages and daily discharges for the period 1917–1922 as estimated by the HOPR. The results of a five-order

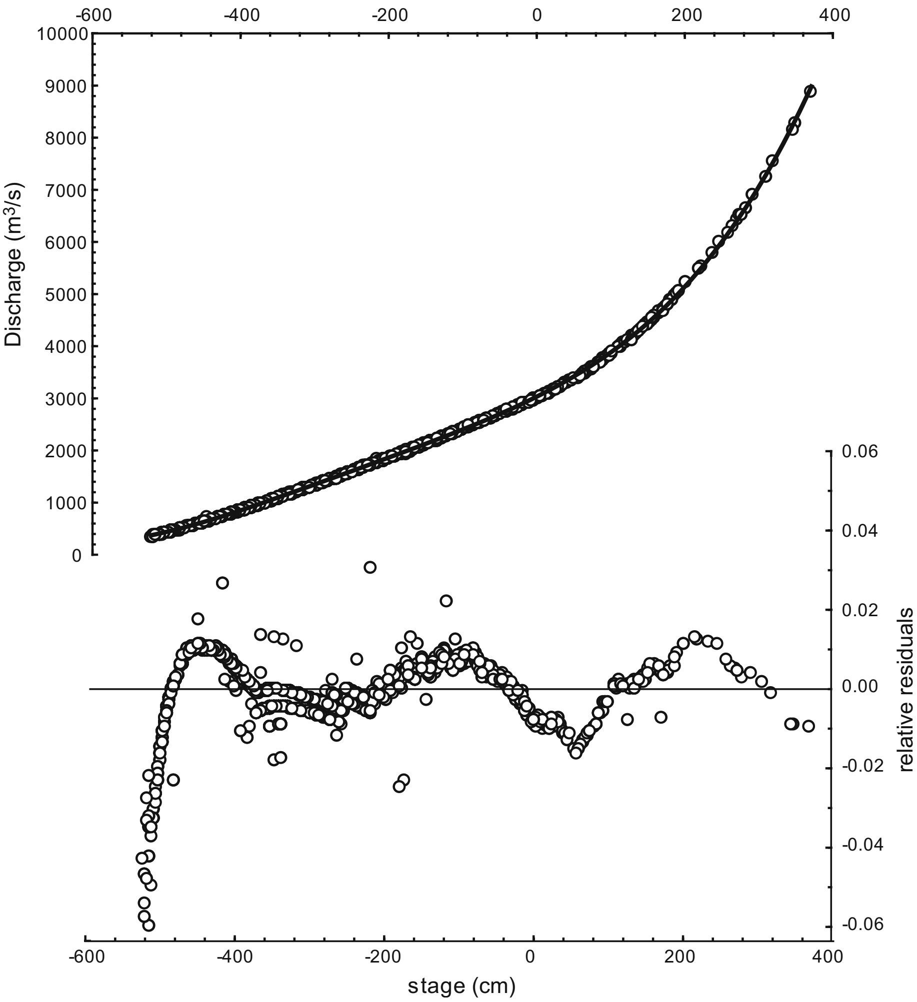


**Fig. 5** *Top panel* scatterplot of daily stages versus daily discharges for the period 1917–1922, with five-order polynomial fit to the data. *Bottom panel* scatterplot of the relative residuals of a five-order polynomial model applied on daily data of discharge (dependent variable) and stage (predictor) for the period 1917–1922

polynomial regression model between stage (predictor) and discharge (dependent variable) data are reported in Tables 5 and 6.[4] The model accuracy is further confirmed by the relative residuals plotted in the bottom panel of Fig. 5, which indicates that model errors are generally below the threshold of 2%, except an apparent overestimation (up to 6%) of very low discharges, i.e., below 500 $m^3/s$. In the light of this evidence, such a regression model is used as an accurate alternative to the empirical DRC established by the HOPR in 1923–1925, the exact equation of which was not found in the available literature.

Figure 6 shows the temporal evolution of the annual-average, -minimum and -maximum daily discharges since 1807, where 1807–1916 data are the results of the stage–discharge conversion described above. The maximum daily discharge observed during the 1807–1916 period is 7,805 $m^3/s$ recorded on 23 October 1872. After 1916, at least seven flooding episodes with daily peak discharge above 8,000 $m^3/s$ occurred, specifically in 1917, 1926, 1928, 1951, 1976, 1994 and 2000, with the absolute maximum discharge observed on 20 May 1926 (9,780 $m^3/s$). The minimum daily discharge observed during the 1807–1916 period was recorded on 12 May 1817 (277 $m^3/s$). After 1917, at least five drought events culminated in a discharge minimum below 300 $m^3/s$: in 1938, 1949, 2003, 2005 and 2006, the latter event coinciding with the minimum daily discharge ever observed (168 $m^3/s$, on 21 July 2006). The statistical significance of the observed apparent sharpening of extremes is assessed by adopting the non-parametric Mann–Kendall (MK) test as statistical procedure for monotone trend detection, which is a common approach in environmental and hydrological studies (e.g., Libiseller and Grimvall 2002; Déry and Wood 2005). MK statistics indicate a clear sharpening of annual-maximum daily discharge (MK = 3.38, $p \sim 0$) and annual-minimum daily discharge (MK = −2.63, $p \sim 0$), whilst changes in the annual-average discharge are apparently negligible since the early nineteenth century (MK = −0.38, $p = 0.71$).

Concerning maximum discharges, a statistically significant shift in the average is found around 1945 (+21%), i.e., slightly before the dramatic flood of 1951. Notably, when the shift component is removed from the time series, the linear-trend component is no longer observed at significant levels (MK = 0.01, $p = 0.98$). Two major facts explain the detected shift.

a) The historical increase in flood peaks at Pontelagoscuro reflects the disruption of the natural and historical functioning of the river caused by the development of infrastructures along the fluvial network to protect urbanized and farming areas in alluvial plains (e.g., Govi and Maraga 2005). In particular, until the late 1950s the defence system was made of discontinuous levee stretches (mostly erected since the late nineteenth century) that allowed a natural damping of the peak-flow discharge of the Upper Po tributaries.
b) The damping effect of many great sub-alpine lakes on high discharges has changed because of lake management, specifically because of the regulation of the water-surface level by sluice-gates. According to Cati (1981), during the flood event of 1917 (when the peak-flow discharge was 8,900 $m^3/s$ and lakes were not regulated) the water-mass storing in sub-alpine lakes, estimated in about $1.6 \times 10^9$ $m^3$, caused an estimated reduction of the peak-flow discharge at Pontelagoscuro of about 400 $m^3/s$. Concerning the flood event of 1951 (when the peak-flow discharge was 11,080 $m^3/s$ and lakes were regulated), an estimated reduction of the peak-flow discharge of at least 430 $m^3/s$ would have been expected if sub-alpine lakes had been unregulated.

[4] The choice of the model is based on best regression results, it has no hydraulic meanings.

**Table 5** Model results of the five-order polynomial regression between daily discharges (dependent) and daily stages (predictor) for the period 1917–1922: summary of the whole model $R$ with statistics of the residuals (included the Kolmogorov–Smirnov $d$ and the Shapiro–Wilk's $W$ tests for normality)

| Whole model | | | | | | |
|---|---|---|---|---|---|---|
| Adjusted $R^2$ | TSS | $F$ ($n_{pred}$,$df$) | Residuals | | | |
| | | | RSS | Var | K–S $d$ ($p$) | $W$ ($p$) |
| 0.999 | 2.054E+9 | $F$(6,2184)=3,012,999 | 297,824 | 136 | 0.11510 (<0.01) | 0.88184 (~0) |

Concerning minimum discharges, shifts are detected only on a CUSUM rank-based test. Apparent shifts are found at 1862 (−20%), 1897 (+25%) and 1938 (−18%), the last shift coinciding with the first drought of the series that characterized the Po River basin during the 1940s.

## 4.3 Analysis of the monthly time series

Figure 7 shows the 1807–2006 monthly time series of Po River discharges. It reproduces the main characteristic observed in the daily patterns of Fig. 6, i.e., an apparent sharpening of extreme values. This apparent behaviour is assessed on a statistical basis by performing MK tests for trend detection on single-month sequences (i.e., data are separately collated according to the belonging month). Partial MK statistics (Libiseller and Grimvall 2002) are also calculated for testing to which degree the MK statistic calculated over a monthly discharge sequence (response series) depends on the presence of trends in the associated precipitation and evapotranspiration sequences (predictor series). According to the MK statistics in Table 7, that concern the period 1831–2003, precipitation has significantly increased in winter whilst changes in the rest of the year are apparently negligible; PET has strongly increased in summer, i.e., when evapotranspiration processes are more significant; discharge values have increased in winter and strongly decreased in summer. According to the partial MK statistics, the increase in discharge values observed in February actually depends on the increase in precipitation; conversely the downward trend observed in summer months does not depend on concurrent changes in precipitation. Downward trends in summer discharges are rather the outcome of an ensemble of changes affecting the annual regime of discharges. Specifically: a weakening of autumn precipitation (although

**Table 6** Model results of the five-order polynomial regression between daily discharges (dependent) and daily stages (predictor) for the period 1917–1922: regression coefficients

| Parameter estimates | | |
|---|---|---|
| Power | Parameter (SE) | $t$ ($p$) |
| 0 | 3,032.572 (0.754) | 4,023.800 (~0) |
| 1 | 7.130 (0.005) | 1,223.117 (~0) |
| 2 | 1.074E-2 (0.004E-2) | 260.014 (~0) |
| 3 | 2.664E-5 (0.001E-5) | 311.672 (~0) |
| 4 | 2.517E-8 (0.001E-8) | 65.537 (~0) |
| 5 | −1.427E-12 (0.001E-12) | −2.369 (0.018) |

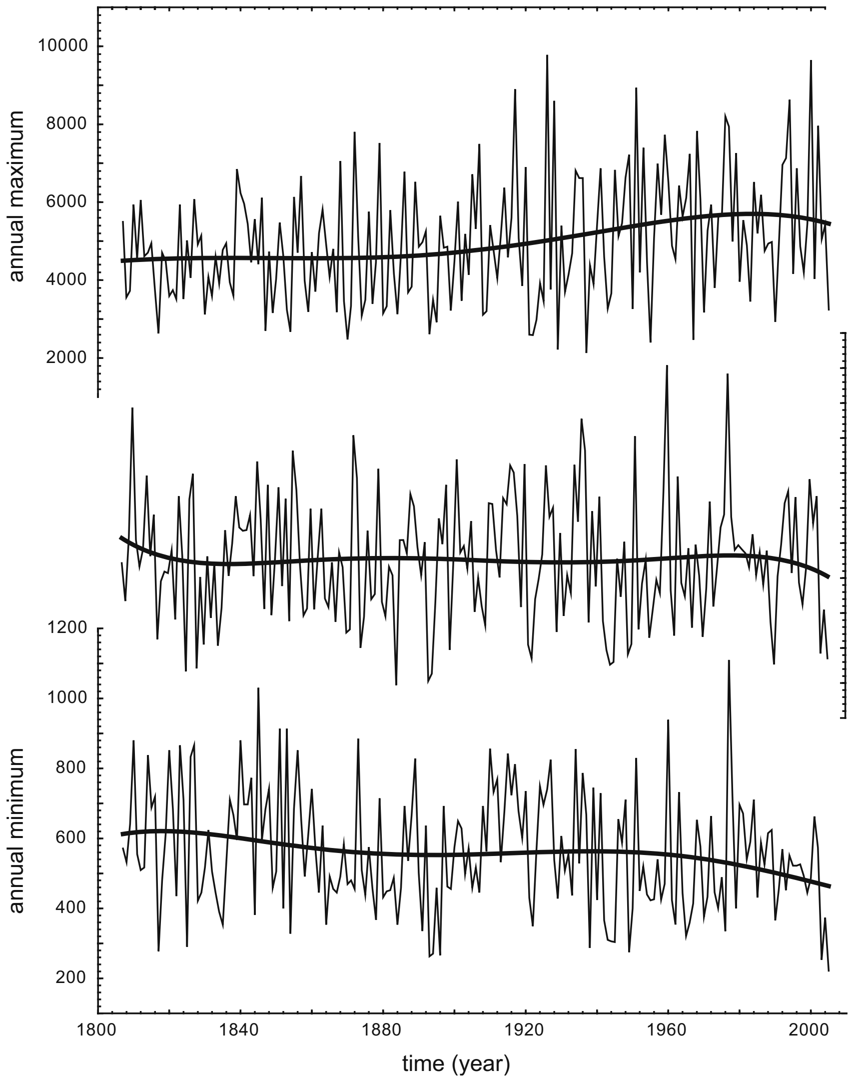


**Fig. 6** Annual maximum (*top*), average (*middle*) and minimum (*bottom*) of daily Po River discharges for the period 1807–2005, with five-order polynomial fit to the data (*thick lines*)

not observed at significant levels); an anticipation of the first annual flood (May instead of June) associated to an anticipation of snowmelt; an increasing demand of water for irrigation originated by strengthening evapotranspiration processes. Therefore, the basin has faced at the same time a lengthening of the summer low-discharge period and an increasing loss of water during runoff processes, which eventually resulted in the observed lowering of summer discharges.

Shift detection on single-month sequences completes the framework for discussing the scenario depicted by the trend statistics. Results of CUSUM tests (results not shown) indicate that a clear downward shift occurred in 1942 in May to August sequences, accounting for an average decrease in late spring/summer discharges of about 23%. An upward shift is detected in May around 1901 (+5%). The synchrony observed between

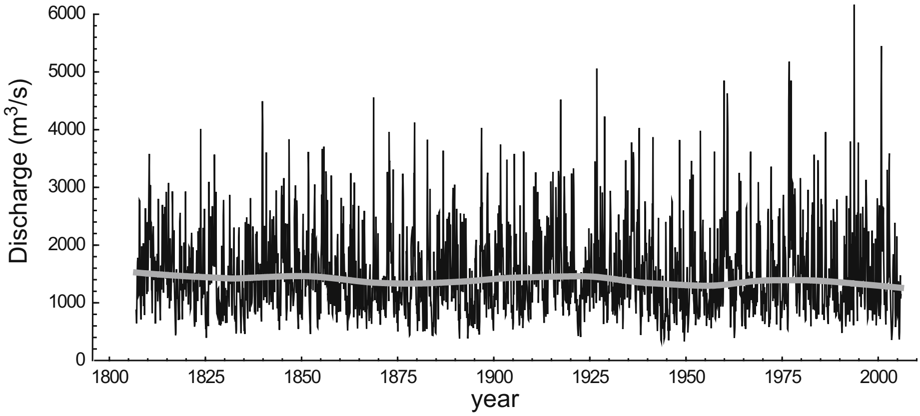


**Fig. 7** Monthly time series of Po River discharges with distance weighted least-square fit to the data

regime shifts in summer discharges and evapotranspiration suggests that the 1940s droughts were the symptom of a sudden climate change, whose effects on the hydrological cycle were possibly worsened by the preceding two-decades of scarce precipitation inflow (i.e., the downward shift detected around 1920.9). In this connection, the anomalous warming and cooling of the Extra-Tropical Pacific which was characterized by the changes in persistent large-scale sea-surface temperature anomalies that occurred approximately on 1921, 1944 and 1973 (Power et al. 1999), suggests that the observed shifts could have a global origin.

In order to avoid the violation of the assumption of independent errors in the non-parametric detection of shifts in discharge data, the deseasoned monthly time series is de-autocorrelated through the equation: $x_t = x_t - \alpha_1 x_{t-1}$, where $\alpha_1 = -0.546$ is the lag-1 autocorrelation coefficient. Shifts in the average are detected (results not shown) above the 5% significant level around 2000.8 (+13%) and 1835.5 (+23%); a likely shift in the *D*–SD is detected around 1992.3 (+47%, $p = 0.08$). Although statistically significant, the shifts in the average are neglected as they may be the result of border effects. The lack of shifts before 1917 further indicates that also the earliest monthly discharge values are likely to be accurately estimated.

The Craddock test of cumulative deviations (Craddock 1979) is commonly adopted in meteorological analyses for checking homogeneity of reconstructed time series. Specifically, the symptom of inhomogeneities due to errors in the reconstructed data is the presence of wide deviations from the optimal zero-value of the Craddock's statistic (e.g.,

**Table 7** MK statistics of monthly sequences of *P*, *E*, (*P*–*E*), discharges (*D*) and discharges conditioned to *P*–*E* (*D* partial)

| | J | F | M | A | M | J | J | A | S | O | N | D |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| *P* | *2.67* | *2.40* | 1.09 | −1.23 | *−2.80* | 0.55 | 0.20 | −1.59 | −1.84 | −1.64 | −0.29 | 1.66 |
| *E* | *5.21* | *2.93* | *4.03* | 0.87 | *2.16* | −1.11 | 1.92 | *3.73* | *2.51* | 0.83 | *2.90* | *4.18* |
| *P*–*E* | *2.43* | *2.05* | 0.62 | −1.35 | *−2.88* | 0.48 | −0.23 | *−2.25* | −1.86 | −1.62 | −0.60 | 1.40 |
| *D* | 1.94 | *2.26* | 1.66 | 0.91 | 0.69 | −1.85 | *−3.81* | *−3.48* | −1.32 | −0.90 | −0.35 | 0.04 |
| *D* partial | 0.62 | 1.24 | 1.28 | 1.60 | *2.19* | *−2.04* | *−3.90* | *−3.14* | −0.53 | −0.16 | −0.25 | −0.93 |

Statistics are calculated over the period 1831–2003 ($N_{cases}=173$). Italicized values indicate the presence of a linear trend in the series significant at $p<0.05$.

Buffoni et al. 1999). Figure 8 shows the Craddock's statistic calculated as cumulative difference between the normalized time series of discharge and precipitation. Deviations are observed between 1880 and the late 1920s (with a minimum around 1910), around 1940 and since the early 1980s.

Figure 8 provides also the cumulative residuals between (*P–E*) and *D*, which is here considered as a semi-quantitative estimate of the $\Delta R$ variable in Eq. 1. According to this approach, catchment reservoirs ($\Delta R$) were characterized by accumulation (i.e., increasing values) between 1880 and 1910, and by an apparent, continuous consumption since 1910. This apparent behaviour would underline the drama of the most recent drought events, that occurred upon a background of deficit comparatively worse than that of the 1940s droughts (1938, 1942–1945 and 1949). The question is whether the observed behaviour is physically plausible or not.

The top panel of Fig. 8 shows the prominent evolutionary features of the annual *E*/*P* ratio and of the discharge autocorrelation ($\alpha_1$). Both these parameters show a behaviour that is apparently consistent with the long-term dynamics associated to $\Delta R$. Concerning the *E*/*P* ratio, a clear upward trend is observed (which means a decreasing percentage of total inflow potentially available for storage) with a local minimum around 1890. The unbalance between the average conditions of PET and precipitation in the pre- and post-1920 shift periods (i.e., pre-: stronger precipitation and lower PET; post-: weaker precipitation and stronger PET) provides further explanation to the temporal evolution of $\Delta R$, specifically to the fact that the amount of water depleted after 1920 is higher than the amount stored in the basin before 1920.

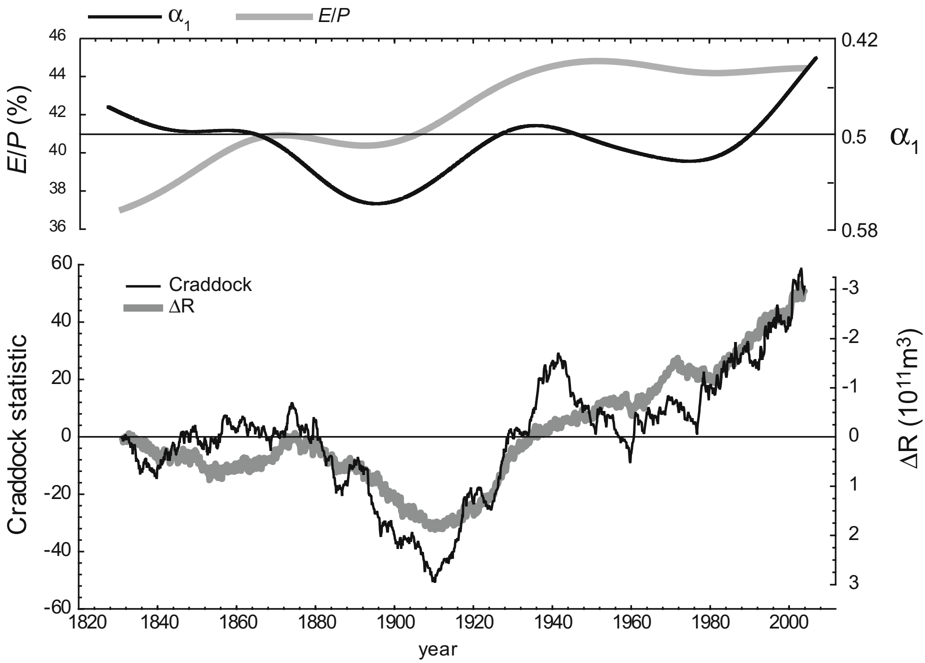


**Fig. 8** *Top panel* lag-1 autocorrelation coefficient ($\alpha_1$) of monthly Po River discharges calculated over a 20-year moving window (axis scale is reversed to facilitate comparison) and estimated annual *E*/*P* ratio (distance weighted least-square fitting). *Bottom panel* Craddock's statistic between monthly Po River discharge and total precipitation over the Po River basin: negative (positive) values indicate precipitation overestimating (underestimating) discharges; dynamics of catchment's reservoirs ($\Delta R$) estimated as cumulative sum of residuals between *P–E* and discharges: positive (negative) values indicate accumulation (consumption)

Concerning discharge autocorrelation, $\alpha_1$-values show a temporary but remarkable increase around 1890 and 1980. An increase in $\alpha_1$-values is associated to a comparatively stronger persistence in monthly discharge signals (i.e., monthly discharges have a more correlated, more skewed but less variable structure), which is amongst the most important property in any hydrologic system as it affects storage capacity of reservoirs, average return periods and drought properties. Highly (not much) autocorrelated discharges may result from more (less) persistent dynamics controlling runoff processes in the basin or rather from more (less) persistent dynamics associated to precipitation.

In order to address to which extent the time-varying persistence of discharges depends on variations in the persistence of precipitation, a comparison is made of peculiar characteristics of the power spectra of precipitation and discharges. In particular, emphasis is devoted to the *reddening* of the discharge power spectrum below the 1-year periodicity, i.e., the gradual decrease in power with increasing frequency observed at high frequencies. According to Milly and Wetherald (2002), a quantitative measures of the redness of a hydrological process $x(t)$ is provided by the normalized ratio $S_X$ of the variance of monthly ($m$) mean anomalies to annual ($a$) mean anomalies ($S_X = \sigma_m^2 / 12\sigma_a^2$), where $\sigma_\tau$ is calculated as the integral of the power spectrum of $x$ from 0 to the frequency associated with $\tau$. For a *white-noise* process (e.g., precipitation) $S$ has a value of unity; a value smaller than unity is indicative of *red-noise* (e.g., discharges).

The time-varying $S$ ratios calculated for precipitation ($S_P$) and discharges ($S_D$) are shown in Fig. 9. As expected, precipitation produces a *white-noise* pattern, with $S_P$ peaks well above unity and of similar magnitude throughout the study period. The $S_D$ pattern shows peaks that are generally far below unity but have much more varying magnitude; in particular, the most recent peaks are apparently approaching unity, which is suggestive of a more impulsive behaviour of discharges during the most recent decades. Despite this, $S_P$

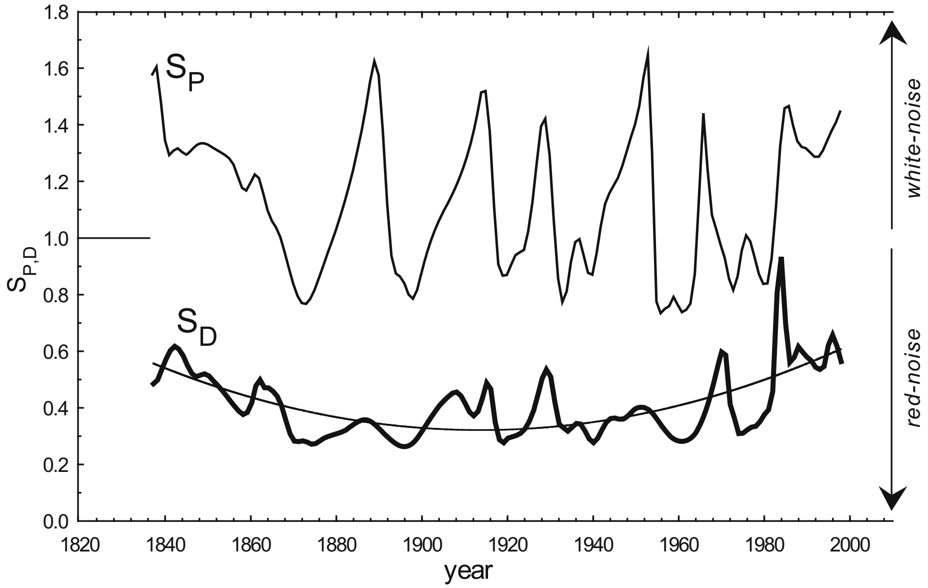


**Fig. 9** Comparison between the normalized ratio ($S$) of the variance of monthly mean anomalies to annual mean anomalies of precipitation ($S_P$) and discharges ($S_D$, with quadratic polynomial fit to the data). $S$ is calculated within a 10-year moving window for the period 1831–2003

and $S_D$ show a remarkable correlation ($r$ = 0.58, $p$ ~ 0, $N$ = 162) which is seen as a synchronic alternation of peaks and valleys. This indicates that signals of varying persistence are apparently conserved in the transformation of precipitation variability to discharge variability. Notably, such periods of strong and weak persistence defines apparent cycles of about 20 years, which is close to the average length of Hale cycles of solar activity (~22 years, e.g., Attolini et al. 1990).

The dampening/*reddening* of original precipitation signals in the discharge spectrum has also interesting spin-offs in the assessment of the statistical significance of low-frequency periodic components. Figure 10 depicts the oscillatory components at periods of 32, 64 and 128 months extracted from the wavelet power spectrum (Torrence and Compo 1998) of monthly precipitation and discharge data. Each spectrum shows when and how much the associated oscillatory mode is manifested in the time series. Two prominent characteristics are observed: (a) the remarkable synchrony at all periods, which further corroborates the hypothesis of a consistent reconstruction of precipitation and discharge data; (b) the similar magnitude of peaks generated by discharges and precipitation at lower (i.e., decadal) frequencies. Notably, the statistical significance of peaks in the spectrum of a given process

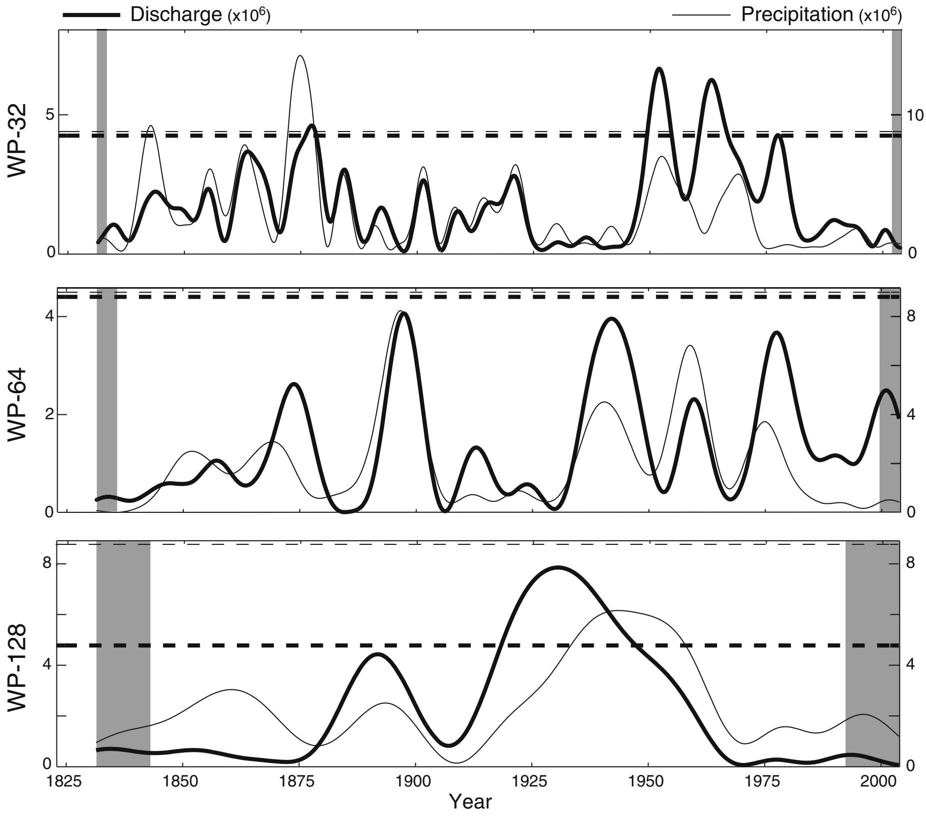


**Fig. 10** Oscillatory components of the wavelet power (WP) spectrum of precipitation and discharges at periods of 32 (*top*), 64 (*middle*) and 128 months (*bottom*) estimated by using the Morlet function as mother wavelet. The *shaded grey* area is the area where edge effects occur. *Dotted lines* indicates the 5% significance level

$x(t)$ can be calculated by multiplying the background spectrum of an appropriate autoregressive process with lag-1 = $\alpha_{1(x)}$ by the 95th percentile value for a chi-square distribution $\chi^2$. This means that the height of the background spectrum of $x(t)$ at a given frequency depends only on the redness of $x(t)$, specifically the height diminishes when $\alpha_{1(x)}$ increases. As a result, peaks of similar magnitude may result below the significance level in the precipitation spectrum but above the significance level in the discharge spectrum, which is observed here, for instance, at the 128-month (~11-year) periodicity. So, Po River discharge is the likely variable to be preferred for assessing the decadal and multi-decadal behavior of the hydrological system at the basin's scale.

The apparently coherent evolution of hydrological processes in the Po River basin provides supporting evidence of a dominant influence of large-scale climate variability (of natural or anthropogenic origin) on the area upon human interventions in the basin such as land-use changes. In this connection, a simple composite analysis with *t* test for statistical significance was performed on precipitation and discharge data by using negative/positive anomalies of the NAO index as grouping criterion. The investigation focuses on period-average values of January–March (JFM) and October–March (OM), i.e., when NAO signals are more stationary. The *t* test results (Table 8) indicate that there exists a robust NAO-related modulation of wintertime hydrological variability, which is observed as stronger (weaker) precipitation and higher (lower) discharges during negative (positive) anomalies of the NAO index. According to these results, when the NAO moves from a strong negative state (NAO < −0.5 SD) to a strong positive state (NAO > 0.5 SD) an average reduction of about 25% and 19% is to be expected, respectively, on OM-average precipitation and discharges. Since a host of studies is increasingly suggesting that enhanced/reduced solar activity can be the precursor of at least some NAO variability (e.g., Boberg and Lundstedt 2002; Thejll et al. 2003; Palamara and Bryant 2004; Lukianova and Alekseev 2005; Zanchettin et al. 2006), accepting the NAO to be amongst the major robust causes of hydrological variability in the area would also imply that the impacts of solar activity must be taken into appropriate consideration when modeling the response of the catchment to different climate and water use scenarios.

**Table 8** Results of composite analyses within January–March (JFM) and October–March (OM) data ($m^3/s$) of precipitation (*P*) and discharges (*D*) grouped under NAO anomalies for the period 1831–2003

| Period | Grouping threshold | Test variable | NAO(+), mean±SD | NAO(−), mean±SD | *t* value (*p*) | Cases NAO(+) | Cases NAO(−) | Variation (%) |
|---|---|---|---|---|---|---|---|---|
| JFM | 0 | *D* | 1,205±390 | 1,407±472 | 3.00 (0.003) | 114 | 59 | −14.4 |
| | | *P* | 1,895±924 | 2,279±930 | 2.58 (0.01) | 114 | 59 | −16.8 |
| | ±0.5 SD | *D* | 1,111±333 | 1,467±431 | 3.60 (<0.001) | 32 | 28 | −24.3 |
| | | *P* | 1,835±793 | 2,496±793 | 2.94 (0.005) | 32 | 28 | −26.5 |
| OM | 0 | *D* | 1,477±522 | 1,550±499 | 0.90 (0.37) | 106 | 66 | −4.7 |
| | | *P* | 2,268±700 | 2,580±782 | 2.72 (0.007) | 106 | 66 | −12.1 |
| | ±0.5 SD | *D* | 1,342±462 | 1,648±523 | 2.39 (0.02) | 35 | 25 | −18.6 |
| | | *P* | 2,116±664 | 2,814±754 | 3.86 (<0.001) | 35 | 25 | −24.8 |

Two thresholds are used to identify positive and negative NAO anomalies: zero and ±0.5 standard deviation (SD). The latter threshold allows to exclude neutral phases of the NAO from the compositing. The expected variation (%) in *P* and *D* when the NAO moves from a negative to a positive state is also indicated.

## 5 Conclusions

This study assessed the viability of backward extending the 1917–present discharge time series of the Po River (Northern Italy) in its closing section at Pontelagoscuro. On this purpose, daily discharge estimates for the period 1807–1916 were calculated from daily stage data collected in archival studies of the Hydrographic Office of the Po River. The extended discharge time series was checked for inhomogeneities through basic trend- and change-point analyses and through comparison with approximated catchment-averages of precipitation and evapotranspiration, which were calculated, respectively, from secular time series of local precipitation (at Turin, Milan and Parma) and local air temperature (at Turin and Milan). Taking into account human interventions on runoff processes, although with unavoidable approximations, the reconstructed data of precipitation, evapotranspiration and discharge provide a coherent picture of the hydrological dynamics in the basin for the 1831–2003 period.

Specifically, the most prominent characteristic of the discharge time series is the sharpening of extremes observed in recent decades. In particular, prolonged drought periods are observed in the 1940s and since 2003, the latter period showing worsening statistics. The likely cause of the lack of prolonged drought periods of similar intensity in the preceding period is thought to be a concomitant change in precipitation (downward shift) and evapotranspiration (upward shift) regimes detected around 1920, which is likely responsible of a progressive depletion of reservoirs in the basin. The increase in peak-flow discharges in recent decades, with values well above the maximum discharge estimated for the nineteenth and early twentieth centuries, is apparently not the result of climate changes but rather the result mainly of the massive levee works along the river network completed in the 1960s.

On decadal and longer time scales, discharge variability is assessed to essentially reflect changes in precipitation patterns, which was observed in particular as concurrent changes in precipitation and discharge persistence and as the presence of peaks of comparable magnitude in the 128-month (i.e., ~11-year) wavelet spectra of precipitation and discharge. Considering that the red-noise background spectrum of discharge is much lower than that of precipitation, river discharge seems the hydrological variable to be preferred for assessing the basin's response to the background climatic variability occurring at the decadal and multi-decadal time scales.

Concerning the seasonal to interannual response to climatic forcing, a robust dependence of wintertime precipitation and discharges on the state of the NAO is assessed on a centennial time scale, this dependence resulting in stronger (weaker) precipitation and higher (lower) discharges during negative (positive) anomalies of the NAO index.

In conclusion, the promising results achieved in this preliminary study encourage further investigations on the extended time series of Po River discharge, notwithstanding partial information about the stage–discharge relationship, which could lead to further reviews for accuracy and, eventually, supplemental corrections on early discharge estimates.

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