Controls of channel morphology and sediment concentration on flow resistance in a large sand-bed river: A case study of the lower Yellow River

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ABSTRACT

Accurate estimation of flow resistance is crucial for flood routing, flow discharge and velocity estimation, and engineering design. Various empirical and semiempirical flow resistance models have been developed during the past century; however, a universal flow resistance model for varying types of rivers has remained difficult to be achieved to date. In this study, hydrometric data sets from six stations in the lower Yellow River during 1958–1959 are used to calibrate three empirical flow resistance models (Eqs. (5)–(7)) and evaluate their predictability. A group of statistical measures have been used to evaluate the goodness of fit of these models, including root mean square error (RMSE), coefficient of determination (CD), the Nash coefficient (NA), mean relative error (MRE), mean symmetry error (MSE), percentage of data with a relative error $\leq 50\%$ and 25% (P_{50} , P_{25}), and percentage of data with overestimated error (POE). Three model selection criterions are also employed to assess the model predictability: Akaike information criterion (AIC), Bayesian information criterion (BIC), and a modified model selection criterion (MSC). The results show that mean flow depth (d) and water surface slope (S) can only explain a small proportion of variance in flow resistance. When channel width (w) and suspended sediment concentration (SSC) are involved, the new model (7) achieves a better performance than the previous ones. The MRE of model (7) is generally $<20\%$, which is apparently better than that reported by previous studies. This model is validated using the data sets from the corresponding stations during 1965–1966, and the results show larger uncertainties than the calibrating model. This probably resulted from the temporal shift of dominant controls caused by channel change resulting from varying flow regime. With the advancements of earth observation techniques, information about channel width, mean flow depth, and suspended sediment concentration can be effectively extracted from multisource satellite images. We expect that the empirical methods developed in this study can be used as an effective surrogate in estimation of flow resistance in the large sand-bed rivers like the lower Yellow River.

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1. Introduction

Flow velocity is a fundamental component of fluvial hydraulic geometry and sediment transport. The three classical flow velocity formulae include the Darcy-Weisbach equation

$$v = \left(\frac{8gRS_f}{f_r} \right)^{1/2}, \quad (1)$$

Chézy equation

$$v = C(RS_f)^{1/2} \quad (2)$$

and Manning equation

$$v = \frac{R^{2/3} S_f^{1/2}}{n} \quad (3)$$

where v is the mean flow velocity; R is the hydraulic radius; S_f is the friction slope (often approximated by water surface slope or channel slope, S); g is the acceleration caused by gravity, and f_r , C , and n are Darcy-Weisbach, Chézy, and Manning flow resistance coefficients, respectively (in SI units).

The central problem in applying these three equations is the estimation of flow resistance coefficients. Quantifying flow resistance can significantly contribute to flood routing, prediction of flow velocity, ecological habitat evaluation, engineering design, geomorphological regime theory development, and other scientific and practical applications (Bathurst, 2002; Ferguson, 2007).

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Investigations on flow resistance have survived for more than a century (Manning, 1891; Keulegan, 1938; Vanoni, 1941, 1946; Vanoni and Brooks, 1957; Qian et al., 1959; Peterson and Mohanty, 1960; Simons and Richardson, 1960; Vanoni and Nomicos, 1960; Simons et al., 1963; Qian and Zhou, 1965; Rouse, 1965; Golubtsov, 1969; Limerinos, 1970; Judd and Peterson, 1969; Bathurst, 1978, 1985, 2002; Bray, 1979; Hey, 1979; Davis and Sutherland, 1980; Griffiths, 1981; Jarret, 1984; Aguirre-Pe and Fuentes, 1990; Bennett, 1995; Dingman and Sharma, 1997; Nikora et al., 1998; Lee and Ferguson, 2002; Ferro, 2003; Ferguson, 2007, 2010; López et al., 2007; Recking et al., 2008; Reid and Hickin, 2008; Comiti et al., 2009; David et al., 2010; Robert, 2011; Ferreira et al., 2012; Nitsche et al., 2012; Schneider et al., 2015). Among these studies, the methods used for quantifying flow resistance can be generally categorized into two major groups: (i) a characteristic particle size approach (or a semiempirical approach) (Keulegan, 1938; Vanoni and Brooks, 1957; Qian et al., 1959; Simons and Richardson, 1960; Vanoni and Nomicos, 1960; Simons et al., 1963; Rouse, 1965; Judd and Peterson, 1969; Bathurst, 1978, 1985, 2002; Bray, 1979; Hey, 1979; Recking et al., 2008; Reid and Hickin, 2008), and (ii) a random field approach (or an empirical approach) (Jarret, 1984; Thorne and Zevenbergen, 1985; Rickenmann, 1994, 1996; Dingman and Sharma, 1997; Nikora et al., 1998; Comiti et al., 2007; Aberle and Smart, 2003; Aberle et al., 2010).

The semiempirical method has been advanced by the success of boundary layer and pipe flow theory proposed by Keulegan (1938), which relates flow resistance to relative submergence in the semilogarithmic form. This method has been successfully applied to predict flow resistance in low-gradient, gravel-bed channels and/or mountain streams, where flow resistance is dominated by grain roughness (Bathurst, 1978, 1985, 2002; Nikora et al., 1998; Lee and Ferguson, 2002; Ferguson, 2007). However, performance has been poor when this method is extrapolated to other streams in which the flow is controlled by different components of flow resistance, such as form roughness or spill roughness (Comiti et al., 2007; David et al., 2010). Thus, empirical approaches have been developed to represent the total flow resistance resulting from different sources of rough elements by which flow resistance (velocity or discharge) can be directly related to reliable, easily measurable channel properties or flow variables by statistical means (Golubtsov, 1969; Riggs, 1976; Jarret, 1984; Dingman and Sharma, 1997; Comiti et al., 2007; David et al., 2010). Among the variables included in the empirical analyses, mean flow depth and/or water surface slope (stream bed slope) are generally considered as the two most important. Some investigations have demonstrated that unit discharge or dimensionless unit discharge exerts an important influence on flow resistance (Rickenmann, 1991; Bjerklie et al., 2005a; Comiti et al., 2007; Ferguson, 2007; David et al., 2010; Rickenmann and Recking, 2011; D'Agostino and Michélini, 2015; Schneider et al., 2015). In fact, unit discharge or dimensionless unit discharge is physically equal to the product of mean flow depth and water surface slope in that it is the function of flow depth and water surface and usually preferable to the use of boundary shear stress for estimating incipient sediment motion in steep streams (Comiti et al., 2007). For streams where bedform roughness is dominant, a random field of bed elevations approach has been proposed (such as standard error of bed elevations and two-dimensional second-order structure function of bed elevations) to characterize flow-dependent sand bed roughness (Nikora et al., 1998; Aberle and Smart, 2003; Aberle et al., 2010).

Because of the complex controls on roughness, large uncertainties have been observed through extensive field observation (Bathurst, 2002; David et al., 2010). Flow resistance is dominant by various controls, which is difficult to completely explain. The existing relationships could not perform reliably over a wide range of flows and hydraulic conditions (Bathurst, 2002; David et al., 2010). Moreover, the uncertainty may result from the varying data sets obtained from different field investigations and flume experiments. The channel form and/or flow variables exhibit large spatial and temporal variability and substantial variability between the investigated sites or experiment design (David

et al., 2010; Ferguson, 2010). A complete understanding of the controls on flow resistance has thus far eluded us.

Flow resistance in sand-bed rivers appears to be more complex than that in gravel-bed rivers, in that sand-bed rivers generally have more easily movable boundary conditions (Vanoni and Brooks, 1957; Simons and Richardson, 1960; Vanoni and Nomicos, 1960). Many efforts have also been devoted to developing flow resistance equations for sand-bed rivers (Vanoni, 1946; Qian et al., 1959; Simons and Richardson, 1960; Simons et al., 1963; Qian and Zhou, 1965; Rouse, 1965; Wang and White, 1993; Wright and Parker, 2004; Yang et al., 2005; Huybrechts et al., 2011; Cheng, 2016). Qian et al. (1959) and Qian and Zhou (1965) argued that roughness was closely related to a given grain size (D_{65} , 65% finer than). Yang et al. (2005) showed that total flow resistance in the large alluvial sand-bed rivers is dominated by sand dune height. The evidence from a large quantity of flume experiments and field investigations also demonstrated the importance of bedforms to flow resistance in sand-bed streams (Vanoni and Brooks, 1957; Simons and Richardson, 1960; Vanoni and Nomicos, 1960; Shen, 1962; van Rijn, 1984; Huybrechts et al., 2011). And various flow resistance models for bedform roughness, which worked independently under different flow conditions, have been proposed (Richardson et al., 1962; Lau, 1983; Wang and White, 1993; Julien and Raslan, 1998; Niemann et al., 2011). The effects of suspended sediment on channel bedform and flow resistance have also been investigated through a series of flume experiments (Vanoni and Nomicos, 1960; Arora et al., 1986; Lyn, 1991). The earlier results showed that the presence of suspended sediment could cause a reduction in flow resistance because of the damping effects of suspended sediment on flow turbulence (Vanoni, 1953; Vanoni and Nomicos, 1960). On the contrary, subsequent flume experiments indicated that the presence of suspended sediment could lead to an increase or a reduction in flow resistance (Arora et al., 1986; Lyn, 1991). Also, bedload transport will tend to increase flow resistance because of the work that the moving water does (Vanoni and Nomicos, 1960; Bjerklie, 2007). In rivers where the Froude number is <1 , flow resistance will be affected by downstream conditions and by cross-channel turbulence and eddy viscosity; thus flow resistance is a three-dimensional characteristic of flow conditions. Therefore, flow resistance in sand-bed rivers is extremely complex because it is a function of flow regime (Richardson et al., 1962; Lau, 1983; Wang and White, 1993; Julien and Raslan, 1998; Niemann et al., 2011), sediment characteristics (Qian et al., 1959; Camenen et al., 2006; Cheng, 2016), and channel geometry (Simons and Richardson, 1960).

The lower Yellow River is a highly transport-limited sand-bed river with excessive suspended sediment concentration (Qian and Zhou, 1965; Wang et al., 2009b; Tables 1 and 2). The river channel bed is composed of fine sand (mean grain size <0.1 mm; Table 2). The high suspended sediment load in the lower Yellow River exerts great influence on channel morphology change (Qian and Zhou, 1965; Wang et al., 2009b; Ma et al., 2012). Although some investigations have studied flow resistance in the lower Yellow River (Qian et al., 1959; Qian and Zhou, 1965; Qin et al., 1995; Cheng et al., 1997; Sun et al., 2005), a clear recognition of the major controls on flow resistance is still lacking. Different from the previous flume experiments, this study aims to identify the major controls on flow resistance through a series of multivariate regression models, and we will discuss their roles in variation of flow resistance. We expect that the results will provide a better understanding of the controls on flow resistance in large sand-bed rivers and will contribute to flood routing in ungauged large sand-bed rivers like the lower Yellow River.

2. Study reaches

The lower Yellow River starts from Mengjin and flows into the sea at Lijin with a length of 768 km (Fig. 1A). The mean annual runoff of the Yellow River is about 37.89 billion m^3 , and the mean annual suspended

Table 1
Basic characteristics of the study reaches at all six stations.

Station	Characteristics of monitoring reach
HK	This reach is wandering-braided mixed. There were river training works on the right bank, and on the left bank is sandy floodplain. The distance between the two artificial levees was about 10 km. The river bed material was fine sand. There often occur intensive scour and deposition. And river channel migrated frequently. During dry season, the flow passed through the main channel. During flood season, flow began to submerge the sand bars near the river banks and became two-thread. River channel width was >3000 m and in the central channel was about 1500 m wide shallow sand bars.
JT	The river channel in this reach was restricted by 8000-m-wide artificial earth levees. River channel is sand bed and wandering-braided. There were two river bends upstream and downstream from the station. During dry season, there were shallow sand bars with some ditches. During flood season, the main channel migrated frequently and severe scour and deposition occurred. The bankfull width is about 3000 m.
GC	This reach was transitional from wandering-braided to meandering. River channel was straight and the cross section is compound. The sand river bed was easily changed with intensive scour and deposition. And the thalweg migrated frequently. On the right bank was about 4200 m wide floodplain.
AS	The reach was relatively straight with sand bed. On the left bank were engineering dikes and on the right bank were hills. River channel was relatively narrow. River pattern of the reach is meandering.
LK	River channel in this reach was relatively straight and river bed is sandy. River channel width was generally <300 m. On the left bank was a 1200-m-wide floodplain. On the right bank was artificial levees built with stones.
LJ	The 2-km-long reach was straight. There was a river bend 800 m upstream from this reach. The river bed was composed of sand and susceptible to scour and deposition. During low flows the thalweg was near the left bank and during moderate and/or high flows moved towards right bank.

sediment load is 0.91 billion tons at Huayuankou during the period of 1950–2009. The lower Yellow River is characterized by heavy sediment-laden flows with average annual suspended sediment concentration of 24.2 kg m^{-3} at Huayuankou station during 1950–2009. The recorded maximum flood was $22,300 \text{ m}^3 \text{ s}^{-1}$ at Huayuankou station in 1958 (Zhao, 1996). The river channel is constrained by artificial levees, and serious deposition often occurs on the floodplains between levees. The lower Yellow River is known as the ‘suspended river’ because the river channel bed is several meters higher than the land surface outside the levees (Xu, 2004). Three reaches are defined based on the longitudinal variation of fluvial geomorphological characteristics: wandering-braided reach, transitional reach, and meandering reach (Fig. 1). Wandering-braided reach is most upstream extending from Mengjin downstream to Gaocun with a length of 299 km. The river channels width at bankfull stage vary from 1 to 3.5 km (Fig. 1B). Mean slope of longitudinal bed profile is about 0.172–0.265‰, and channel sinuosity is about 1.15. The height between river channel bed and bank surface is generally <2 m. The river channel is quite unstable and easy to migrate laterally. From Gaocun to Taochengpu, the river channel is transitional from wandering-braided to meandering (Fig. 1C). The length of this reach is about 165 km, and river channel at bankfull stage varies from 0.5 to 1.6 km. The mean bed slope is about 0.148‰, and channel sinuosity is >1.3. The height from river channel bed to bank surface ranges from 2 to 3 m. The river channel in this reach is more stable than in the wandering-braided reach. The meandering reach is between Taochengpu and Ninghai (Fig. 1D). The river channel has width varying from 0.3 to 0.8 km. The mean river bed slope is 0.1‰, and channel sinuosity is 1.2. The height from river channel bed to bank surface is generally >3 m. Owing to the constraints by artificial levees, river channel in this reach remains stable (Pan et al., 2006).

Sanmenxia Dam, the first large dam in the Yellow River basin, was put into use in 1960 (Fig. 1A). Before the Sanmenxia Dam was put into use, flow processes in the lower Yellow River had received less disturbance. The large amount of sediment deposited in the river channel resulted in shrinkage of river channel size. During 1965–1973, the reservoir was regulated to store low flow with low suspended sediment concentration during the dry season and to release turbid flows with high suspended sediment concentration during the flood season. The high sediment load during the flood season exceeded flow transport capacity and caused severe river channel deposition (Ma et al., 2012).

The hydrometric stations used in this study include Huayuankou (HK), Jiahetan (JT), Gaocun (GC), Aishan (AS), Luokou (LK), and Lijin (LJ) (Fig. 1A). These six gauging sites belong to different river reaches (Table 1). Stations HK and JT are located in the wandering-braided reach; GC in the transitional reach; and AS, LK, and LJ in the meandering reach.

3. Data

In the Yellow River basin totally, >300 hydrometric stations have been built by the Yellow River Conservancy Commission since the 1950s. Data sets collected at the gauging stations were issued as ‘Hydrological Year Book’ for the use of engineering design and scientific research after standard calibration and compilation. Regular measurements of hydrological variables (including water stage, discharge, water width, water depth, velocity and suspended sediment concentration) comply with national standards published by the Hydrological Bureau of Ministry of Water Conservancy and Electric Power (1962). The procedures for hydrometry, sediment sampling, and laboratory analyses as well as the calibrating methods were

Table 2
The range of hydraulic variables in flow measurement in the study reaches^a.

Station	Q ($\text{m}^3 \text{ s}^{-1}$)	Q _b (max.) ($\text{m}^3 \text{ s}^{-1}$)	w (m)	d (m)	v (m s^{-1})	S (10^{-4})	SSC (kg m^{-3})	n ($\text{m}^{-1/3} \text{ s}$)	N
HK	70–7323	7520	286–2360	0.35–3.04	0.32–3.11	0.2–5.8	1.73–244	0.005–0.044	98
	83–6452	6900	176–2820	0.58–2.15	0.45–2.57	0.4–7.5	0.956–143	0.006–0.312	86
JT	72–7122	7575	112–1710	0.58–2.15	0.37–3.29	0.14–4.2	0.638–202	0.002–0.027	92
	81–5010	5210	226–1400	0.56–2.55	0.502–3.5	0.563–5.92	0.77–140	0.006–0.030	99
GC	72–7729	7630	248–6565	0.72–2.23	0.33–3.15	0.1–4.9	1.09–193	0.003–0.029	115
	60–4254	4540	171–2090	0.58–2.16	0.14–2.56	0.2–2.7	0.942–132	0.003–0.111	65
AS	71–6036	6300	229–439	0.62–6	0.42–3.56	0.37–2.4	0.562–156	0.005–0.024	121
	–	5900	–	–	–	–	–	–	–
LK	77–6440	6920	113–855	1.13–10.1	0.31–3.1	0.32–7.6	0.365–130	0.008–0.041	105
	89–5052	5290	83–288	1.45–6.9	0.405–2.62	0.16–6.7	0.89–93.9	0.006–0.054	91
LJ	23–6559	6870	123–557	0.96–5.7	0.19–2.84	0.1–2.8	0.087–142	0.004–0.034	129
	47–5027	5250	53–550	0.85–4.5	0.303–2.92	0.1–2.1	0.137–135	0.006–0.027	99

^a Note: N is the number of measured data. Upper lines: 1958–1959, lower lines: 1965–1966

Fig. 1. Remote sensing images showing the Yellow River basin and the study reaches. SMX: Sanmenxia Dam, XLD: Xiaolangdi Dam, MJ: Mengjin County, HK: Huayankou, JT: Jiahetan, GC: Gaocun, AS: Aishan, LK: Luokou, LJ: Lijin, TP: Taochengpu, NH: Ninghai. The image for the Yellow River basin was produced from MODIS (August 2004) (A) and the images showing the different reaches from Landsat 8 using band 6–5–4 (B–D) (The acquisition data of Landsat 8 near HK, GC, and LK are, respectively, 23 April 2015, 18 May 2015, 25 April 2015). (B) The wandering-braided reach near HK, (C) the transitional reach near GC, and (D) the meandering reach near LK.

basically consistent with the international criterions (Boiten, 2003). All hydrological variables (including sediment samples) mentioned above were generally sampled at least 25 times a year. During large flood events, the measurements were repeated >5 times every day. During each measurement, >25 verticals were measured in a cross section and 2–5 points at each vertical. Then the cross-sectional average was calculated. Discharge is calculated using the velocity-area method based on daily velocity measurements. Systematic errors existed in the measurement of total sediment discharge at hydrometric stations because of the difficulty in measurements of suspended sediment load near bed and bedload. However, the relative error in the sediment concentration measurement was <1% (Wang et al., 2009b). It was estimated that bedload only accounted for 0.3–0.7% of the total sediment load (Long and Zhang, 2002). The measured suspended sediment loads at some hydrometric stations were underestimated ~3–7% (Li and Long, 1994). By contrast, water surface slope was regularly measured only during several early years. At least six water level gauging sites were built at both banks along a 30 km reach. When mean flow depth, mean cross section velocity, and water surface slope were eventually collected and calculated, the Manning roughness coefficient (n) were derived using Eq. (3). The data sets measured during two different periods

(1958–1959, 1965–1966) were chosen because of their adequacy and completeness. During both periods, the value of water surface slope was measured over different flow events >90 times. During 1958–1959 the lower Yellow River was in a quasi-natural condition with less human disturbance; while during 1965–1966 the river was considerably influenced by the operation of the Sanmenxia Dam. The hydrologic variables used in this study include discharge (Q), river channel width (w), mean flow depth (d), mean flow velocity (v), water surface slope (S), suspended sediment concentration (SSC), and Manning roughness (n) (Table 2). The selected measurements represent a wide range of flow conditions below bankfull discharge. The Q , SSC, and n vary by three orders of magnitude and w , d , v , and S by two (Table 2).

4. Methods

4.1. The transfer between $(8/f_r)^{0.5}$ and n

The flow resistance is expressed in the form of $(8/f_r)^{0.5}$ in order to present a visual comparison with the results from previous investigations.

According to flow Eqs. (1)–(3), $(8/f_r)^{0.5}$ can be related to the other two flow resistance coefficients C and n :

$$(8/f_r)^{0.5} = \frac{C}{g^{1/2}} = \frac{d^{1/6}}{ng^{1/2}} \quad (4)$$

where d is the mean flow depth. For the reaches in this study, w is much larger than d , so the hydraulic radius can be replaced by d . Therefore, given Manning coefficient n , f_r can be derived by Eq. (4). Manning coefficient n is inversely proportional to $(8/f_r)^{0.5}$.

4.2. Proposed models

As mean flow depth (d) and water surface slope (S) are the two defining variables in flow continuity equations, they have routinely been involved in the development of flow resistance models (Golubtsov, 1969; Riggs, 1976; Bray, 1979; Jarret, 1984; Coon, 1998; Dingman and Sharma, 1997; Bjerklie et al., 2003; David et al., 2010). Hence, most of the traditional flow resistance models can be written as

$$(8/f_r)^{0.5} = a_1 d^{b_1} S^{c_1}. \quad (5)$$

Then Jarret's (1984) formula can be reformulated as $(8/f_r)^{0.5} = d^{0.33} S^{-0.38}$. We find that $(8/f_r)^{0.5}$ is positively correlated with d and negatively correlated with S .

In fact, the variables included in model (5) can not explain all the variance of total flow resistance, which necessitates inclusion of other important controls. Channel width and suspended sediment concentration, which play an important role in river channel adjustment of the lower Yellow River, are adopted into the multiple regression models. Then, we get two new models as follows:

$$(8/f_r)^{0.5} = a_2 d^{b_2} S^{c_2} w^{d_2} \quad (6)$$

$$(8/f_r)^{0.5} = a_3 d^{b_3} S^{c_3} w^{d_3} (SSC)^{e_3}. \quad (7)$$

The three models (5)–(7) are applied to the six stations in the lower Yellow River, respectively. Before multiple regression analysis is performed, the models are transformed into their logarithmic forms. Then the coefficients and exponents in the models are derived by the least square procedure, which has the advantage of minimizing the errors of the data set, especially those with large magnitude (López et al., 2007). The statistical parameters are calculated using the antilogarithmic version of the value derived from the multiple regressions. The statistical results will make it possible to select the best model for predicting flow resistance in different reaches.

4.3. Statistical measures

An array of statistical parameters are used to evaluate the goodness-of-fit of the three calibrating models, including root

mean square error (*RMSE*), coefficient of determination (*CD*), the Nash coefficient (*NA*), mean relative error (*MRE*), mean symmetry error (*MSE*), percentage of data with a relative error $\leq 50\%$ and 25% (P_{50} , P_{25}), percentage of data with overestimated error (*POE*) (Nash and Sutcliffe, 1970; Bjerklie et al., 2005a; López et al., 2007).

The *RMSE* is the root mean square error between observed and predicted data,

$$RMSE = \sqrt{\frac{\sum_{i=1}^N (P_i - O_i)^2}{N-1}} \quad (8)$$

where P_i is the predicted i th value of the variable, O_i is the observed i th value of the variable, and N is the number of data.

The *CD* is the square of Pearson's product-moment correlation coefficient and can be interpreted as the proportion of the total variance in the observed data that can be explained by the model. It has been calculated as

$$CD = \left(\frac{\sum_{i=1}^N (O_i - \bar{O})(P_i - \bar{P})}{\sqrt{\sum_{i=1}^N (O_i - \bar{O})^2 \sum_{i=1}^N (P_i - \bar{P})^2}} \right)^2 \quad (9)$$

The *NA* is defined as (Nash and Sutcliffe, 1970)

$$NA = 1 - \frac{\sum_{i=1}^N (P_i - O_i)^2}{\sum_{i=1}^N (O_i - \bar{O})^2} \quad (10)$$

The Nash coefficient evaluates the agreement between predicted and observed values. A value of $NA = 1$ indicates a perfect agreement between predicted and observed values.

The *MRE* is the mean relative error, namely the ratio of the absolute error between prediction and observation to the observed value

$$MRE = \frac{100}{N} \sum_{i=1}^N \frac{|P_i - O_i|}{O_i} \quad (11)$$

The *MSE* statistic is calculated as

$$MSE = \frac{100}{N} \sum_{i=1}^N \frac{P_i - O_i}{O_i} \quad (12)$$

This statistical parameter is used to depict whether the predicted value is overestimated or underestimated. If $MSE = 0$, it means that

Table 3

At-a-station relationships of velocity and Manning roughness at all six stations.

Station	1958–1959								1965–1966							
	F_r	$v = kQ^m$			$n = \alpha Q^\beta$			R^2	F_r	$v = kQ^m$			$n = \alpha Q^\beta$			R^2
		k	m	R^2	α	β	R^2			k	m	R^2	α	β	R^2	
HK	0.106–0.820	0.078	0.359	0.756	0.034	−0.162	0.156	0.138–0.616	0.152	0.275	0.495	0.054	−0.197	0.285		
JT	0.126–0.691	0.075	0.413	0.823	0.096	−0.343	0.587	0.162–0.717	0.064	0.442	0.870	0.024	−0.106	0.124		
GC	0.124–0.730	0.071	0.420	0.894	0.033	−0.189	0.333	0.040–0.638	0.036	0.493	0.695	0.089	−0.293	0.205		
AS	0.115–0.595	0.048	0.500	0.917	0.040	−0.200	0.381	–	–	–	–	–	–	–		
LK	0.083–0.457	0.050	0.478	0.931	0.015	0.056	0.039	0.103–0.378	0.048	0.485	0.919	0.021	−0.015	0.001		
LJ	0.061–0.505	0.033	0.521	0.955	0.073	−0.230	0.441	0.105–0.514	0.044	0.479	0.818	0.052	−0.206	0.239		

Fig. 2. The relationships between flow resistance (f_r) and six controlling variables: (A) mean flow depth (d), (B) water surface slope (S), (C) river channel width (w), (D) suspended sediment concentration (SSC), (E) ratio of river channel width to mean flow depth (w/d), and (F) flow discharge (Q).

predicted value agreed perfectly with observed value on the whole. If $MSE > 0$, the predicted value has an overestimation, otherwise it has an underestimation.

Furthermore, in order to identify the best model, three model selection criteria—Akaike information criterion (AIC), Bayesian information criterion (BIC) (López et al., 2007), and modified model selection

Table 4

The power law relationship between flow resistance and four controlling variables.

Station	d		S		w		SSC	
	Exponent	R^2	Exponent	R^2	Exponent	R^2	Exponent	R^2
HK	−0.26	0.0123	0.5052	0.1314	−0.531	0.0895	−0.438	0.2328
JT	−1.175	0.1894	1.1723	0.3658	−0.872	0.2412	−0.707	0.5558
GC	−0.933	0.0964	0.1955	0.0198	−0.872	0.3279	−0.42	0.3825
AS	−0.817	0.2353	−0.438	0.0319	−4.041	0.4937	−0.457	0.5196
LK	0.3574	0.0809	0.2349	0.1107	−0.579	0.0559	−0.048	0.0134
LJ	−1.053	0.1651	−0.651	0.1336	−1.907	0.4817	−0.418	0.5636

Table 5
Multiple regression statistics from calibration data (1958–1959) for models (5)–(7).

Station	$(8/f_r)^{0.5} = a_1 d^{b_1} S^{c_1}$				$(8/f_r)^{0.5} = a_2 d^{b_2} S^{c_2} W^{d_2}$					$(8/f_r)^{0.5} = a_3 d^{b_3} S^{c_3} W^{d_3} (SSC)^{e_3}$					
	a_1	b_1	c_1	$R^2(adj.)$	a_2	b_2	c_2	d_2	$R^2(adj.)$	a_3	b_3	c_3	d_3	e_3	$R^2(adj.)$
HK	2.978	0.207	−0.268	0.155	0.105	0.104	−0.351	0.387	0.313	0.179	−0.360	−0.374	0.135	0.333	0.548
JT	0.527	0.370	−0.465	0.409	0.038	0.317	−0.473	0.387	0.611	0.124	−0.184	−0.507	0.041	0.307	0.761
GC	3.994	0.586	−0.239	0.159	0.278	0.260	−0.311	0.311	0.330	0.145	−0.247	−0.527	0.001	0.370	0.657
AS	16.503	0.332	−0.065	0.166	2.2E-6	−0.203	−0.156	2.610	0.463	5.7E-4	−0.343	−0.190	1.528	0.206	0.596
LK	9.408	−0.095	−0.092	0.133	0.093	−0.130	−0.206	0.677	0.382	0.147	−0.381	−0.356	0.297	0.242	0.706
LJ	103.021	0.287	0.183	0.153	0.009	−0.111	−0.135	1.142	0.475	0.289	−0.373	−0.176	0.432	0.230	0.639

criterion (MSC) from AIC by Koeppenastrop and DeCarlo (1993) and Bjerklie et al. (2005a)—are also used in this study.

$$AIC = N \cdot \ln \frac{\sum_{i=1}^N (P_i - O_i)^2}{N - k - 1} + 2(k + 1) \quad (13)$$

$$BIC = N \cdot \ln \frac{\sum_{i=1}^N (P_i - O_i)^2}{N - k - 1} + k \cdot \ln(n) \quad (14)$$

$$MSC = \ln \left(\frac{\sum_{i=1}^N (O_i - \bar{O})^2}{\sum_{i=1}^N (P_i - O_i)^2} - \frac{2k}{N} \right) \quad (15)$$

where k is the number of parameters in the model. The AIC is among the most popular and versatile strategies for model selection, and in essence it is a penalised version of the attained maximum log-likelihood for each model (Akaike, 1974; Claeskens and Hjort, 2008). This model selection criterion was developed to deal with the trade-off between the goodness-of-fit and complex of a model (Akaike, 1974; Burnham and Anderson, 2002). However, it does not provide any assurance about model quality in an absolute sense. Although Bayesian information criterion (BIC) is closely related to AIC, BIC is also adopted here because the penalty term is larger in BIC than in AIC (Claeskens and Hjort, 2008). These indices are used to identify the best model when more than one model is compared with the same database. The less the value of AIC and BIC, the more precise the predicted value is. On the contrary, the model with the largest MSC value is considered as the most appropriate one for describing the observed data.

5. Results

5.1. Variation of channel roughness with varying discharge

At-a-station hydraulic geometry provides simple and direct characterization of relationships between channel morphology and hydraulic factors (Leopold and Maddock, 1953a). Because velocity is a function

of roughness, the rate of mean velocity with varying discharge reflects the variation of roughness of channel boundaries (López et al., 2007). The relationships between mean flow velocity, channel roughness and discharge are analyzed respectively during the two periods (1958–1959, 1965–1966) (Table 3).

The exponent and correlation coefficient show a strong positive relationship between mean flow velocity (v) and flow discharge (Q) at all six stations (Table 3). During 1958–1959, exponent m ranges from 0.359 to 0.521, with an average of 0.45 for all stations. During 1965–1966, exponent m varies from 0.275 to 0.493, with a mean value of 0.43. We find that average m is almost the same during these two periods, which is similar with the result derived by Leopold and Maddock (1953a).

However, channel roughness generally shows a negative correlation with discharge, which is consistent with some previous observations (Vanoni, 1941; Leopold and Maddock, 1953b; Comiti et al., 2007). Exceptionally, the channel roughness at Luokou Station exhibits nonmonotonic changes (Fig. 2F). Hence, the way to extrapolate flow resistance by the relation between roughness and flow discharge is not always reliable and precisely.

The higher correlation coefficient (R^2) between velocity and discharge shows that flow discharge performs a better explanation to velocity than to roughness (Comiti et al., 2007; Rickenmann and Recking, 2011). We argue that the flow resistance in the lower Yellow River is controlled by other factors, besides mean flow depth (d) and water surface slope (S). The reasons will be discussed in detail in the following sections.

Froude number (F_r ; here can be defined as $v/\sqrt{gd}$) is used to characterize the instantaneous flow regime. The results (Table 3) show that all the measured flow events from all the stations are in the lower regime (subcritical) ($F_r < 1$). The measured maximum F_r is generally higher in wandering-braided and transitional reaches than that in meandering reaches. The maximum F_r was captured at discharges much lower than maximum estimated bankfull discharge except at Gaocun Station. The maximum F_r at Gaocun Station occurred at about $7700 \text{ m}^3 \text{ s}^{-1}$, during the flood season in 1958.

Although good correlation exists between mean flow velocity and flow discharge, it is difficult to popularize this method in flow estimation and flood routing because the difficulty in data collection and estimation of flow discharge, especially in the large sand-bed rivers such as the lower Yellow River.

Table 6
Significance level of the exponents of different variables in models (5)–(7).

Station	$(8/f_r)^{0.5} = a_1 d^{b_1} S^{c_1}$			$(8/f_r)^{0.5} = a_2 d^{b_2} S^{c_2} W^{d_2}$				$(8/f_r)^{0.5} = a_3 d^{b_3} S^{c_3} W^{d_3} (SSC)^{e_3}$				
	a_1	b_1	c_1	a_2	b_2	c_2	d_2	a_3	b_3	c_3	d_3	e_3
HK	0.05	0.02	<0.01	<0.01	0.23	<0.01	<0.01	0.01	<0.01	<0.01	0.06	<0.01
JT	0.32	<0.01	<0.01	<0.01	<0.01	<0.01	<0.01	<0.01	0.04	<0.01	0.51	<0.01
GC	0.01	<0.01	<0.01	0.06	0.06	<0.01	<0.01	<0.01	0.02	<0.01	0.98	<0.01
AS	<0.01	<0.01	0.49	<0.01	0.01	0.04	<0.01	<0.01	<0.01	<0.01	<0.01	<0.01
LK	<0.01	0.22	0.04	<0.01	0.04	<0.01	<0.01	<0.01	<0.01	<0.01	<0.01	<0.01
LJ	<0.01	0.01	0.03	<0.01	0.27	0.07	<0.01	0.27	<0.01	<0.01	<0.01	<0.01

Table 7
Statistical results for calibration of model (5).

Station	RMSE	CD	NA	MRE (%)	MSE (%)	P_{50} (%)	P_{25} (%)	POE (%)	AIC	BIC	MSC
HK	11.110	0.219	0.195	33.410	9.174	82.65	56.12	39.80	483.022	488.777	0.166
JT	14.434	0.644	0.606	32.345	8.487	85.86	53.26	45.65	502.292	507.857	0.906
GC	13.460	0.198	0.174	31.507	7.694	82.09	62.61	37.39	609.005	615.240	0.148
AS	11.617	0.160	0.115	27.996	3.918	87.60	58.68	38.84	604.568	610.955	0.078
LK	5.829	0.120	0.104	20.515	2.973	92.38	74.29	48.57	381.269	387.231	0.057
LJ	10.702	0.109	0.073	30.481	5.799	81.40	54.26	49.61	622.641	629.221	0.032

5.2. Calibration of the proposed models

Before the three multivariable power-law models are calibrated, the contribution of each individual variable to flow resistance is examined by the simple power-law relationship. The results show that flow resistance (f_r) changes with mean flow depth (d) and water surface slope (S) with greater scatter than channel width (w) and suspended sediment concentration (SSC) (Fig. 2), and thus d and S explain less of the variability of flow resistance. This necessitates the inclusion of w and SSC into the development of multivariable models (Table 4).

The results of multiple regressions for the three calibrating models (Eqs. (5)–(7)) are shown in Tables 5 and 6. For model (5), except at Luokou and Lijin stations, although the exponent of S at Aishan is not significant at 0.05 (Table 6), $(8/f_r)^{0.5}$ is positively correlated with d and negatively correlated with S , which agrees with the finding of Jarret (1984). The distinction at Luokou station can probably be attributed to the large uncertainty of the relation between channel roughness (n) and flow discharge (Q) as shown in Fig. 2F. The RMSE at six stations ranges from 5.829 to 14.434, with an average of 11.192. The model (5) has ~30% of mean relative error (MRE) and 6% overestimation (MSE), although POE is generally <50%. >85% of the measurements have predicting errors within 50%, and almost 60% within 25%. However, the low values for adjusted correlation coefficients (R^2) derived from model (5) imply that d and S explain a small proportion of variance in $(8/f_r)^{0.5}$ (Table 7). And CD and NA indicate that model (5) has poor performance for predictability of flow resistance (Table 7). We conclude that there are other variables that produce additional flow resistance in the lower Yellow River. Hence, when w and SSC are included in models (6) and (7), the two new models have a significantly improved adjusted R^2 .

In model (6), $(8/f_r)^{0.5}$ has a negative correlation with S and a positive correlation with w at all six stations (Table 5). The rate of change of S with $(8/f_r)^{0.5}$ varies between -0.135 and -0.473 , while that of w ranges from 0.311 to 2.61, which has a larger variability than that of S . However, the relation between $(8/f_r)^{0.5}$ and d varies between different stations. The $(8/f_r)^{0.5}$ is positively correlated with d at Huayuankou, Jiahetan, and Gaocun, and negatively correlated at Aishan, Luokou, and Lijin (Table 5). Meanwhile, the exponent of d is not significant at 0.05 at Huayuankou, Gaocun and Lijin (Table 6). The CD and NA show that model (6) has a better predictability than model (5) (Table 8). The averages of RMSE and MRE in model (6) are 9.3% and 22.7%, respectively. A slight overestimation (3.52%) occurs despite POE is also <50%.

Model (6) produces <50% errors for 90% of the measurements and <25% errors for 70% of the measurements.

Model (7) provides the best regression and statistical results than models (5) and (6) (Tables 5–9). Model (7) has the highest adjusted R^2 , CD, and NA compared with the other two models, which show that SSC explains a large proportion of variability of flow resistance in the lower Yellow River. Consistently, at all six stations, $(8/f_r)^{0.5}$ is negatively correlated with d and S , and positively correlated with w and SSC (Table 5). In model (7), the exponents of w at Jiahetan and Gaocun indicate that the variation of w does not make much of a contribution to flow resistance (Table 5), and it is not statistically significant at Jiahetan and Gaocun (Table 6). The averaged RMSE and MRE derived from model (7) are 7.5% and 17.3%, respectively. The average overestimation at the six stations is about 2.42%, and the percentage between overestimation and underestimation (POE) is very approximate (Table 9).

Based on the above analyses, model (7) has the best performance. The MRE derived from model (7) is generally <20%, which is relatively lower than those in previous studies (Bathurst, 2002; López et al., 2007; David et al., 2010). Moreover, the three model selection criteria consistently exhibit that model (7) is the best calibrating model for flow resistance in the lower Yellow River.

In order to visually compare the predictability for the full range of flow resistance derived from model (7), predicted $(8/f_r)^{0.5}$ against observed $(8/f_r)^{0.5}$ is plotted in Fig. 3. At Huayuankou, Jiahetan, Gaocun, and Aishan, when $(8/f_r)^{0.5}$ is less than a given threshold, it is apparently overestimated at which the corresponding flow discharge is in the range of low flows ($<900 \text{ m}^3 \text{ s}^{-1}$) (Table 10). This could be related to the intensive variation of channel bedform at low flows when the flows have enough power to erode the river bed and produce ripples and dunes. Comparatively, Luokou station $(8/f_r)^{0.5}$ is overestimated and Lijin station underestimated in the range of all flow discharges (Table 10).

5.3. Validation of model (7)

Here, we adopt a new data sets collected during 1965–1966 to validate model (7). To show the goodness of fit of validation of model (7), the statistical measures adopted in calibration procedure are used here (Table 11). At the other four stations except at Luokou, all the statistics do not exhibit good performance of model (7) (Table 11), and Fig. 4 shows that model (7) generally performs poorly in the lower Yellow River. This may suggest that the coefficients derived from the calibration data sets are not suitable for the validation data sets that come from a different period. The controls on flow resistance in the

Table 8
Statistical results for calibration of model (6).

Station	RMSE	CD	NA	MRE (%)	MSE (%)	P_{50} (%)	P_{25} (%)	POE (%)	AIC	BIC	MSC
HK	10.210	0.344	0.320	28.420	7.130	88.78	63.27	46.94	469.508	477.848	0.329
JT	11.337	0.776	0.757	24.488	4.025	89.13	64.13	41.30	460.906	468.993	1.394
GC	11.686	0.428	0.378	25.180	2.322	88.70	65.21	37.39	579.528	588.508	0.430
AS	8.859	0.500	0.485	20.263	2.828	90.91	75.21	42.98	542.021	551.204	0.630
LK	4.720	0.453	0.413	16.189	1.114	96.19	82.86	48.57	339.996	348.612	0.486
LJ	9.011	0.359	0.343	21.796	3.705	92.25	66.67	47.29	581.305	590.744	0.378

Table 9
Statistical results for calibration of model (7).

Station	RMSE	CD	NA	MRE (%)	MSE (%)	P ₅₀ (%)	P ₂₅ (%)	POE (%)	AIC	BIC	MSC
HK	8.175	0.578	0.564	22.333	4.619	93.88	65.31	46.94	428.997	439.921	0.785
JT	8.613	0.862	0.860	18.047	2.069	95.65	80.43	45.65	413.394	424.003	1.950
GC	8.512	0.678	0.670	17.360	1.628	95.65	77.39	43.48	509.698	521.422	1.079
AS	8.365	0.552	0.541	19.120	2.204	95.87	79.34	42.98	531.158	543.137	0.741
LK	3.177	0.741	0.734	12.203	1.077	98.10	89.52	47.62	259.929	271.199	1.298
LJ	8.138	0.470	0.464	17.339	2.939	96.90	79.85	56.59	558.053	570.352	0.581

lower Yellow River should be complex, and the primary controls may shift with varying flow regime and sediment input especially under the regulation of Sanmenxia Dam. The detailed interpretation will be given in the following section.

6. Discussion

6.1. Controls on flow resistance

As the two fundamental characterizations of flow velocity, mean flow depth (d) and water surface slope (S) have been considered as the most fundamental variables in flow resistance prediction and/or flow discharge estimation (Golubtsov, 1969; Riggs, 1976; Jarret, 1984; Dingman and Sharma, 1997; Bjerklie et al., 2003, 2005b). However, our results show that mean flow depth (d) and water surface slope (S) could only explain a smaller proportion of the variance in flow resistance in the lower Yellow River (Fig. 2; Tables 4 and 5). On the contrary, river channel width (w) and suspended sediment concentration (SSC) have stronger explanatory power. Therefore, with the inclusion of these two variables, models (6) and (7) could explain the greater proportion of variability in flow resistance (Tables 5–9). Model (5) has a poor performance probably in that the flow discharge and sediment transport rate are no unique functions of d and S (Vanoni and Brooks, 1957). A given slope could be produced from several equilibrium flow conditions (Kennedy, 1961). Apparently, flow velocity increases with

the rise of d , w , SSC, and thus Q , while Froude number (F_r) increases with the rise of SSC and Q (Fig. 5). However, the close relationship of the decrease in flow resistance (f_r) with the increase of F_r (especially in high F_r) suggested complex controls on flow resistance at high flows, because the F_r , which works as a good indicator of flow regime and energy expenditure, is dominated by various factors. The statistical results and model selection criterions for models (6) and (7) indicate the significance of channel morphology (w/d , ratio of river channel width to mean flow depth) and suspended sediment concentration (SSC) in flow resistance in the lower Yellow River. On the one hand, SSC impose significant impacts on changing river channel morphology (Leopold and Maddock, 1953b; Xu, 2004; Fig. 7). The w/d generally reduced with the increase in Q and SSC (Figs. 6A and 7A). The high floods with high suspended sediment concentration tend to change river channel from wide-shallow to narrow-deep (Wang et al., 2009b; Fig. 6B–F), and a narrow-deep river channel is relatively flat with less high protrudes as the wide-shallow channel (Fig. 6B–F). The transition of river bedform from ripple-sand dune at low flows to flat bed at high concentrated flows has been demonstrated by the flume experiments (Simons and Richardson, 1960; Vanoni and Nomicos, 1960; Richardson et al., 1962). The increase in w/d means the reduction on flow resistance (Fig. 2E). On the other hand, SSC influences flow resistance by damping turbulence and concomitant reduction of turbulence diffusion coefficients (Vanoni and Brooks, 1957). High concentrated flow events could cause high sediment delivery, which will facilitate

Fig. 3. Predicted $(8/f_r)^{0.5}$ plotted against observed $(8/f_r)^{0.5}$ for the calibration datasets during 1958–1959 at all six stations in the lower Yellow River using Eq. (7). The three lines respectively correspond to perfect agreement and $\pm 25\%$ error.

Table 10

The statistical summary for overestimated or underestimated flow resistance and discharge^a.

Station	$(8/f_r)^{0.5}$	Q ($m^3 s^{-1}$)
HK	• 7.19–19.93	70.10–837 (6922)
JT	• 11.43–19.47	71.52–684.85
GC	• 12.04–28.79	72.32–614.70
AS	• 15.05–29.10	71.20–825.35
LK	• 9.76–14.57	77.29–6439.86
LJ	• 40.27–85.88	448.72–5188.37

^a Note: At Lijin station, the range of underestimated $(8/f_r)^{0.5}$ and the range of corresponding discharges are reported, and at the other stations the range of overestimated $(8/f_r)^{0.5}$ and the range of corresponding discharges are reported.

Table 11

Statistical results for validation of model (7).

Station	RMSE	CD	NA	MRE (%)	MSE (%)	P_{50} (%)	P_{25} (%)	POE (%)
HK	10.388	0.334	−0.559	39.663	28.713	74.71	49.43	67.82
JT	7.282	0.285	0.154	19.751	−0.554	91.92	75.76	37.37
GC	20.479	0.452	0.013	93.969	90.860	70.77	47.69	83.08
AS	—	—	—	—	—	—	—	—
LK	6.061	0.958	0.937	15.815	−6.746	98.90	78.02	31.87
LJ	11.426	0.388	−0.022	22.906	−16.947	96.97	60.61	20.20

the suppression of flow turbulence and reduction of sediment shear stress and riverbed drag (Wang and Liang, 2000).

Special attention should be focused on the complex changes of flow resistance (f_r) with flow discharge (Fig. 2F) at Luokou Station. Because velocity increases with the rise of d , w , and thus Q through a power law relationship (Table 3), the nonmonotonic variation of flow

resistance suggests more complex controls on flow resistance during different flow events than flow velocity (Fig. 5). For discharge $<1500 m^3 s^{-1}$, the flow resistance shows a decreasing trend because the river bedform is dominated by ripple and dunes at low flows. When flow discharge is $>1500 m^3 s^{-1}$, the flow resistance increases slightly with discharge. This may arise from the energy expenditure for excessive sediment transport during high concentrated flows (Wang and Liang, 2000).

As mentioned in the introduction, bed grain size plays an important role in flow resistance especially for gravel-bed rivers or mountain streams because the relative submergence (the ratio of mean depth to grain size) is relatively small (<10 ; Bathurst, 1985, 2002). However, in the lower Yellow River, the relative submergence in the lower Yellow River is quite large (10^4 – 10^6) (Huybrechts et al., 2011); and thus the effects of individual sediment particle on channel roughness can be neglected. Hence, we can draw that the flow resistance in the lower Yellow River is dominated by river bedform roughness resulting from high concentrated flows.

The fluvial processes have been changes by a series of water and sediment regulation activities resulting from climatic change and human activities in the long term (Williams and Wolman, 1984; Lambeck and Chappell, 2001; Gray et al., 2002; Osterkamp and Gray, 2003; Ma et al., 2012). The large percent reduction in gradient of the lower costal-plain reach is inferred to be largely the result of the reduced flows and sediment deposition caused by dam impoundment and water diversion and partly from the increase in sea level by climatic change during the Holocene (Osterkamp, 1978; Gray et al., 2002; Osterkamp and Gray, 2003; Lambeck and Chappell, 2001). The gentler slope in the lower Yellow River is indicative of lowered flow velocity and flow resistance (Fig. 2B; Osterkamp and Gray, 2003). However, the variation of flow resistance is controlled by complex controls as shown in model (7), which could not completely be explained by each individual variable. More work is needed to further assess the variation

Fig. 4. Predicted $(8/f_r)^{0.5}$ plotted against observed $(8/f_r)^{0.5}$ for the validation datasets during 1965–1966 at the five stations in the lower Yellow River using Eq. (7). The three lines respectively correspond to perfect agreement and $\pm 25\%$ relative error.

Fig. 5. The relationships between different flow-resistance (f_r) related variables. (A) The relationship between mean flow depth (d) and mean flow velocity (v), (B) the relationship between channel width (w) and mean flow velocity (v), (C) the relationship between suspended sediment concentration (SSC) and mean flow velocity (v), (D) the relationship between Froude number (Fr) and flow resistance (f_r), (E) the relationship between flow discharge (Q) and Froude number (Fr), and (F) the relationship between suspended sediment concentration (SSC) and Froude number (Fr).

of flow resistance resulting from the human-induced changes in flow regime and sediment flux in the lower Yellow River.

6.2. Spatial and temporal variations of flow resistance

The data set used in the precedent studies was generally divided into two random parts, one for calibration and the other for validation (Dingman and Sharma, 1997; Bjerklie et al., 2005a; López et al., 2007). The data used for validation may include similar information with that used for calibration. Hence, López et al. (2007) used the same data set for calibration and validation through a cross validation method that randomly divided the data set into two equal parts. The cross validation

can eliminate the uncertainty of the model to a large extent. Although many established models could have good performance for the investigating sites, they usually fail when applied to new sites because the controlling factors on flow resistance in different types of rivers and their contribution to flow resistance are quite different (Fig. 2F; Bathurst, 2002). In this study, the data set used for validation was collected from the same gauging station, but from different periods with different flow conditions. The poor performance of the calibrating model in validation shows that even if at the same site, if the flow regime and sediment input vary, flow resistance will be quite different. The substantial variations substantially limit the application of flow resistance models. Consequently, future work should focus on further

Fig. 6. The 30 river channel cross sections measured during different flow events at Huayuankou station in 1959. The temporal variations during each measurement of cross sections are shown. The frequent measurements during flood season could reflect the river channel shift from wide-shallow to narrow-deep. The river changes generally from multichannel at low flows to one main channel at high flows.

investigating the spatiotemporal variations of controls on flow resistance and establishing physical-based models for flow resistance in the lower Yellow River.

6.3. Potential use of remote sensing in the development of flow resistance models of the lower Yellow River

Large-scale modelling and/or quantifying of flood events in areas without ground-based gauging sites will necessitate the use of a flow resistance model combined with remotely measured or estimated hydraulic variables and/or channel hydraulic properties (Smith et al.,

1996, Smith and Pavelsky, 2008; Bjerklie et al., 2005a; Bjerklie, 2007). Remote sensing data from varying sources and sensors with high temporal resolution could significantly contribute to retrieve useful information of different landscape units. Many efforts have been devoted to develop efficient algorithms for acquiring hydraulic geometry variables and suspended sediment concentration from various available satellite images, which has been demonstrated to be practical and reliable (Smith et al., 1995, 1996, Smith and Pavelsky, 2008; Bjerklie et al., 2003, 2005b; Xu et al., 2004; LeFavour and Alsdorf, 2005; Leon et al., 2006; Wang et al., 2009a; Birkinshaw et al., 2010; Durand et al., 2010; Miller et al., 2014; Legleiter, 2015). For example, effective river channel

Fig. 7. (A) The relationship between the ratio of river channel width to mean flow depth (w/d) and flow discharge (Q). These data are retrieved from each measurement of the cross sections at Huayankou station in 1959. (B) The relationship between river channel elevation change and the sediment transport coefficient. The sediment coefficient is defined as the ratio of suspended sediment concentration to flow discharge (Ma et al., 2012), which could better reflect the relative amount of sediment input to flow discharge.

width can be derived from different categories of remote sensing images, such as ASTER, LANDSAT, SPOT, ERS, QuickBird, and MODIS (Smith et al., 1995, 1996; Smith and Pavelsky, 2008; Bjerklie et al., 2003, 2005b; Xu et al., 2004; Birkinshaw et al., 2010), and water surface stage can be retrieved from the microwave remote sensing image data (SAR, LiDAR) and digital elevation model (DEM) (Bjerklie et al., 2003, 2005b; LeFavour and Alsdorf, 2005). The SSC can be retrieved from different bands of MODIS, Landsat ETM+ and MSS, TM, and AVHRR based on the water body's spectral characteristics by relating spectral reflectance in the visible and near infrared bands to SSC (Ritchie and Cooper, 1988; Chen and Curran, 1992; Islam et al., 2001; Doxaran et al., 2002; Miller and McKee, 2004; Wang et al., 2009a). In terms of mean flow depth, rational methods have been developed that are based on the observations of zero flows (Leon et al., 2006) and/or river channel geometry (Miller et al., 2014; Legleiter, 2015) from satellite altimeter data. Therefore, with the rapid advancement of remote sensing techniques, the empirical methods developed in this study can be popularized in numerical modelling for flood routing and water resources management in the large rivers like the lower Yellow River.

7. Conclusions

Accurate estimation of flow resistance is pivotal to flood routing, flow discharge and velocity estimation, river restoration and management, and water resources management. A number of empirical and semiempirical models for predicting flow resistance have been

developed during the past century, and the subsequent examinations demonstrated that they didn't have good performance because of the spatially varying flow conditions. In this study, three empirical models are proposed to predict flow resistance in the lower Yellow River using hydrometric data sets from six stations during 1958–1959. A group of statistical measures are used to evaluate their goodness-of-fit, including root mean square error (RMSE), coefficient of determination (CD), the Nash coefficient (NA), mean relative error (MRE), mean symmetry error (MSE), percentage of data with a relative error $\leq 50\%$ and 25% (P_{50} , P_{25}), and percentage of data with overestimated error (POE). Meanwhile, three model selection criteria are calculated to identify the best model, including Akaike information criterion (AIC), Bayesian information criterion (BIC), and a modified model selection criterion (MSC). The results show that mean flow depth and water surface slope can only explain a small proportion of variance in flow resistance. When river channel width and suspended sediment concentration are considered, all the statistics indicate that this model has the best predictability. The MRE from this model is generally $<20\%$, which is relatively better than that reported by previous studies (Bathurst, 1985, 2002; David et al., 2010). The best calibrating model is validated using the data from the same stations during 1965–1966. The statistical results show that this model produces large uncertainty. This may suggest that even if the controls on flow resistance are the same, the coefficients and exponents in the models derived from different data sets may be significantly different. Consequently, when one perfectly established model is applied to a new data set, great

uncertainty will arise. However, with the rapid advancements of earth observation techniques, information about channel width, mean flow depth, and suspended sediment concentration can be effectively retrieved from multisource remote sensing sensors. Hence, the empirical methods developed in this study can be popularized in numerical modelling for flood routing and water resources management in ungauged areas with the help of remote sensing techniques.

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