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Fluvial Processes and Their Future Magnitudes: Combined Field Observation and Simulation Approaches

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Contents

1. Introduction

The concept of uniformitarianism that "the present is the key to the past" has been applied in the studies of past rate and intensity variability of processes controlling erosion, transport and deposition (Orme, 2002; Eriksson et al., 2005). Stream channels evolve continuously, and the Davisian theories of river erosion cycles and stream adjustment to the geomorphological structure have been widely shown and accepted for over a century (Morisawa, 1989). The same physical laws and mechanics of fluvial processes have been assumed to prevail in the past; therefore they can be assumed to prevail also in the future. Hydrological, hydraulic, and sediment transport processes are all interlinked in open channel flow, and connection between physical and environmental understanding of river processes and evolution are needed for developing fluvial studies (Biswas, 1970; Richards et al., 1987).

Sediment particles move, when the fluid forces (here defined also as flow characteristics), such as velocity, shear stress and stream power, are greater than the resisting forces of particles and river bed (Hjulström, 1935; Bagnold 1977; Baker & Ritter, 1975; Knighton, 1984; Bagnold, 1986; Baker & Costa, 1987; van Rijn, 2007). Changes in discharge cause changes in these fluid forces (Blum & Törnqvist, 2000). These in turn cause river channels to change, and the magnitude changes of future fluvial processes can be said to depend on future discharges (Lane et al. 2007). In this present study, the concept of magnitude of fluvial processes is defined as the forces excerted on the river channel, as well as consequent fluvial transport and channel modifications during different discharges. Thus, it also includes the magnitudes of the discharges themselves. The role of floods in the erosion and deposition of stream channels has been a controversial topic in fluvial studies, as major floods have been observed to only produce either little channel change, highly temporary effects or great modifications (Richards, 1986; Baker, 1988). Thus, need exists for studies of the effects of different magnitude discharge events on channel evolution and upon the erosion-sedimentation potential in different riverine environments.

Although the empirical studies of hydrology and fluvial processes have developed over the centuries (Orme, 2002), there are still occasions, such as future and infrequent high discharge events, when it is impossible to perform field measurements. For example great floods have sometimes been analysed after their occurrence and by applying model simulations (Tinkler & Wohl, 1998). These simulations enable the prediction of the real world behaviour under a set of naturally occurring circumstances (Beven, 1989). Numerical simulations are performed with models that attempt to describe the physical system at different scales (Raudkivi, 1979). Models are simplifications of reality and it has been stated that in complex natural systems, such as rivers, there will always be levels of detail that cannot be simulated (Beven, 1989; Coulthard et al., 2007). Recently, in addition to the physical models performed in a laboratory flume environment, numerical models have nevertheless become a widely available tool for fluvial studies, since they can provide spatial and temporal insights into fluvial processes and geomorphology (Richards, 1986; Ingham & Ma, 2005). In the present work, such simulation approaches include hydrological, hydrodynamic (a synonym for hydraulic models that depict the river hydraulics) and morphodynamic simulations. Moreover, channel evolution can be simulated by applying morphodynamic (sometimes also called morphological) models, which include both river bed elevation changes, as well as sediment transport by flowing water.

With the awareness of possible future climate change there has been a parallel increase in the national and regional scale hydrological simulations of climate change impacts on discharges (e.g. Vehviläinen & Huttunen, 1997; Andréasson et al., 2004; Lehner et al. 2006; Graham et al., 2007; Beldring et al., 2008; Steele-Dunne et al., 2008), while the use of hydrological models has become a standard tool (Bergström, 2001). The climate change signal is transferred from different climate scenarios (i.e. the combination of climate model and emission scenario) into hydrological models. According to the IPCC (2007), the observed widespread warming of the atmosphere suggests that it is unlikely that the global climate change of the past 50 years is explainable without external forcing and very likely that it is not due to natural causes alone. Global warming may intensify the global water cycle, affect temperature and rainfall and consequently, affect both river flows and flood risk (Arnell & Reynard, 1996; Milly et al., 2002; Maaskant et al., 2009). Future warming is expected to be greatest over land and in northern latitudes (IPCC, 2007). Moreover, based on different emission scenarios, the range of global sea level rise is predicted to be 0.18–0.59 m by 2100 (IPCC, 2007). However, this rate of sea level rise may not be linear, since the relative sea level is predicted to accelerate after the middle of the 21st century (Johansson et al., 2004).

According to the Directive of the European parliament and of the council on the assessment and management of flood risks (European commission, 2007), the impacts of climate change on the occurrence of floods should be taken into account when assessing flood risks. Flood risk can be defined as the combination of the physical characteristics of the flood event (the hazard) and its potential consequences, such as the actual exposure to floods and the coping capacity (de Moel et al., 2009). Demands are posed for general evaluations of the changes of flood discharges, flood inundation areas and possible flood hazard, due to climate change in different parts of Europe (European commission, 2007). Both short and long term estimates are needed for planning, since the changes in floods may not be linear. Further, the geomorphological processes in flood risk management should also be taken into account, particularly where aggradation occurs (Lane et al., 2007).

Future hydro- and morphodynamic simulations can be performed by applying future discharges simulated using hydrological models as the boundary conditions. In recent years, those studies where numerical models have been used for future suspended sediment transport and channel evolution have started to appear (e.g. Lane et al., 2007; Thodsen et al., 2008; Verhaar et al., 2008; Ganju & Schoellhamer, 2009; Gomez et al., 2009; Verhaar et al., 2010; Verhaar et al., 2011). However, these future fluvial processes have been analysed to a lesser extent by detecting changes in the flow characteristics (velocity, shear stress and stream power) and thus, in erosion and sedimentation potential also. According to Verhaar et al. (2011), what is needed are two-dimensional, long term unsteady flow simulations of bed load transport, which also take bank erosion into account.

There is also a lack of studies, where the climate change signal from multiple scenarios has been included in simulations of future flood inundation. To date, it has been mostly only the effects of present or past different magnitude discharge events on flood inundation that have been analysed (Horrit & Bates 2002; Bates et al., 2005; Néelz et al., 2006; Pappenberger et al., 2006; Hunter et al. 2007; McMillan & Brasington 2007; Koivumäki et al., 2010). Although a European-scale flood risk analysis based on past floods has been performed (i.e. without numerical simulation) (Schmidt-Thomé et al., 2006), there also exists a need for further local scale and future change analyses. While Thonon & Klok (2007) have found that future flood frequencies increase together with inundation, they did not perform an actual hydrodynamic simulation for the inundation extent. Further, in many studies, the effects of the present and future sea level rise on the extent of inundation have been performed without hydrodynamic simulation (Gambolati et al., 2002; Li et al., 2006). Even though some numerical simulations of coastal inundation and backwater effects due to sea level do exist (Lamb et al., 2012; Parker et al., 2008; Purvis et al., 2008), more analysis is needed on the combined effects of discharge and sea water levels on inundation and the fluid forces exerted on river channel.

The focus of this study lies in improving the scientific knowledge of fluvial processes and their future magnitudes in eastern Fennoscandia (i.e. selected locations from Finland and Finnmark, northern Norway) (Fig. 1). Several climate scenarios and hydrological simulation results are applied in numerical hydro- and morphodynamic simulations and supported by extensive field observation campaigns in the period 2007–2010. A national scale assessment of future run-off is produced for Finland (paper I) and case studies are conducted in two rivers representing different hydrological and geomorphological conditions: A) the subarctic, sandy and braided lower Tana River (northern Norway, papers II and III), B) lower Kokemäenjoki River (coastal area of south-west Finland, papers I and IV), whose channel consists mainly of cohesive sediments. Comparisons are made between these different river reaches.

The main objectives of the thesis are:

1) To derive more knowledge of the present fluvial transport and channel evolution during different magnitude discharges in different rivers (papers III and IV).

2) To estimate the possible future discharge changes in different rivers and seasons on a century time-scale (papers I, II and IV).

3) To study the future changes in flow characteristics, such as velocity, shear stress and stream power, within and between different river reaches due to possible discharge changes (papers II and IV).

4) To estimate the combined effects of discharge and sea level changes on flood inundation and flow characteristics (papers I and IV).

5) To improve our understanding of the effects of possible discharge and sea level changes on morphodynamics in different rivers (papers I–IV).

Paper I presents the national scale study of the climate change effects on flood discharges and inundation extents in Finland for the periods 2010–2039 and 2070–2099. Since the European-scale assessments (Lehner et al. 2006; Dankers & Feyen 2008) have been contradictory and unreliable on a detailed scale, the overall picture of flood hazard changes in Finland with consistent methods and scenarios has also been missing. This study presents a unique approach, where the effects of 20 climate scenarios on flood discharges were analysed at 67 sites. These climate scenarios were based on combinations of different global and regional climate models (GCM and RCM) and emission scenarios. The simulations were performed with a conceptual hydrological model, the Watershed Simulation and Forecasting System (WSFS) from the Finnish Environment Institute. The estimation of possible changes in runoff and floods during different seasons was included in the study (season boundary changes were not considered). In addition, their effects were combined into the flood inundation investigation at four specific study sites, which exhibited different watershed types and hydrological conditions. A two-dimensional (2-D) hydrodynamic model (TUFLOW)

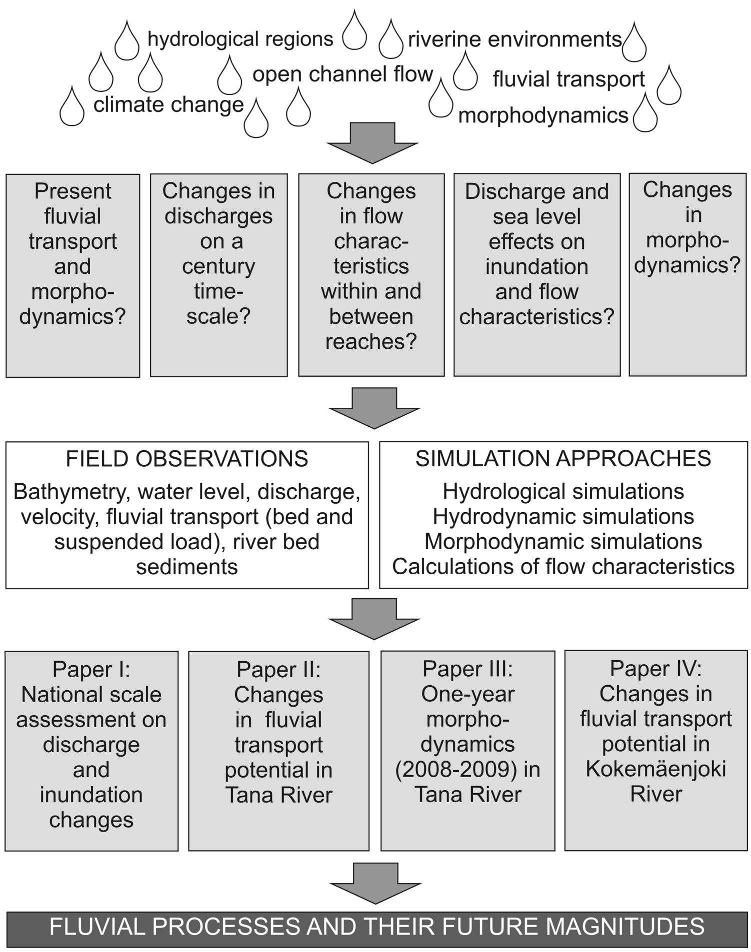

Fig. 1. The research approaches for this thesis of fluvial processes and their future magnitudes.

was applied in a flood inundation analysis of 100-year discharges (i.e. recurrence interval of once per hundred years).

In paper II, the effects of climate change on the future discharges and the spatial variability of velocity, bed shear stress and stream power per unit area were detected in the subarctic lower Tana River. The variation in spatial erosion power between the present and future frequent (HQ1/2a or 2-year, i.e. high discharge event with a recurrence interval of every other year) and infrequent discharges (HQ 1/250a or 250-year) were simulated with the 2-D hydrodynamic model (TUFLOW). Future discharge scenarios (2070–2099) were simulated with WSFS and using climate scenarios combined from different global and regional climate models and emission scenarios. Although previous studies of climate change impacts on hydrology and discharges have been done in some subarctic catchments of Fennoscandia, less emphasis has been given to the possible changes in flow characteristics (velocities, bed shear stresses and stream powers per unit area). The relative changes in flow characteristics were compared between the frequent and infrequent discharges. Based on different climate scenarios, the flow magnitude changes in different seasons were also analysed (season boundary changes were not considered) and their role in sediment transport compared to the present situation. This study contributes to the future studies of northern areas, which have been found important, since they are expected to be one of the most sensitive environments to climate change (IPCC, 2007).

Paper III represents the application of a morphodynamic simulation in a large braided river system within a one year time span (2008–2009), including two spring high discharge (1- and 2-year) events. There has been discussion as to whether the cumulative effect of annual low discharges could be greater than the effect of frequent floods over a longer time span in different environments (Leopold et al. 1964; Baker 1977). Thus, more information was sought concerning temporal and spatial variations in channel evolution in the lower Tana River during periods of different magnitude discharges using the two-dimensional morphodynamic model (TUFLOW MORPH), as well as field measurements, in particular bathymetric data. The relative importance of different magnitude discharge events on river bed and braid evolution were detected. The fulfilment of these analyses required testing the model's capability to simulate the morphodynamics of the broad braided reach. A multidimensional morphodynamic modelling approach has not previously been performed for a one-year time span in a large subarctic sandy braided river, where the highest discharges occur in the snow melt period. Moreover, the simulations have also normally been performed only for relatively short discharge events. Therefore, this paper also contributes to the studies of fieldwork- and simulation-based morphodynamics.

As a comparison to the subarctic river environment, paper IV presents a study of the impacts of future discharge changes and different sea water level situations on the erosion and sedimentation potential along the lowest ~40 km of the Kokemäenjoki River in south western Finland. Although detection was performed for different seasons (season boundary changes were not considered), it was also undertaken during infrequent and frequent discharge events. It has been forecast that due to climate change, discharges, sea level and erosion intensity and therefore, fluvial sediment transport also, will all change by the end of 21st century (Andréasson et al., 2004; Johansson et al., 2004; Thodsen et al., 2008; Verhaar et al., 2010). Following the analysis of the present river bed sediments, as well as changes in suspended transport and bed elevation changes, the daily boundary shear stresses and their changes in different seasons were simulated for the periods 1971–2000 and 2070–2099 with a 1-D hydrodynamic model (HEC-RAS). Four different discharge scenarios were applied, based on two different climate signal transfer methods, different global and regional climate models and emission scenarios. Since different sea level conditions were also employed in these simulations, it enabled the analysis of the combined effects of discharge and sea level on flow characteristics along the longitudinal river profile. The spatial variability of erosion and sedimentation potential along the river reach was detected by analysing the bed shear stresses of the average and infrequent 100-year discharge events against the measured bed sediment properties, such as their critical bed shear stress thresholds. More knowledge was also gained concerning the seasonal magnitude changes in the fluvial processes.

2 Theoretical background of fluvial studies

2.1 Fluvial processes

2.1.1 Development of process understanding and measurements

Studies on fluvial processes have been performed for centuries, since the roots of hydrology, including measurements of open channel flow, date back to antiquity. Although the concept of discharge calculation by multiplying the cross-sectional area by the velocity was first defined during ancient times, it was only generally applied in seventeenth century, when the first systematic river bed configuration studies and an understanding of the principles of continuity had also already emerged (Biswas, 1970; Orme, 2002). Since the seventeenth century, not only quantitative hydrology has evolved but also the concepts of the hydrological cycle and continuity equation. The hydrological cycle is a closed system and therefore, it conforms to the principle of conservation of mass (Freeze & Harlan, 1969). Further, according to the principle of momentum, the rate of change in the momentum flux for a selected control volume equals the sum of the forces acting on the control volume (Chanson, 1999). The understanding of fluvial geomorphology emerged concurrently, as did the comprehension of flow velocity variations with depth and slope, stream flow acceleration and opposing bed resistance, while the relationship between channel geometry and sediment scour and fill developed also (Orme, 2002).

Since the studies of experimental physics in the eighteenth century, open channel hydraulics and morphodynamics and thus, the present computational fluid dynamics (CFD) have developed (Biswas, 1970). It is important at this point to mention Du Buat (1734–1809), who was one of the first researchers to study both theoretically and experimentally the effect of flow on a loose sediment bed and who developed the shear-resistance concept (Gomez, 1991; van Rijn, 1993). The consciousness of the importance of the bed material composition, particularly the grain sizes, on erosion and on the relationship between the initiation of motion and velocity emerged later, at the beginning of 20th century. Among the first pioneering works to represent erosion and sedimentation thresholds for different sized particles were those by Schoklitsch (1914), Hjulström (1935) and Shields (1936). Although Gilbert & Murphy (1914) had already attempted to develop a relationship between velocity, depth, slope and discharge, the estimations of sediment transport based on the relationships between bed material sediments and bed shear stress or stream power became common only since the 1960's (Laursen, 1958; Bagnold, 1966, 1977, 1986; Leopold, 1980; Williams, 1983; Baker & Costa 1987). According to Leopold (1980), three important concepts concerning fluvial processes have developed since the work of Gilbert & Murphy (1914): 1) there is a relationship between flow characteristics (velocity, shear stress and stream power) and channel material size, 2) the width, depth and velocity vary together with discharges in cross-sections, and 3) various dependent factors adjust river channels toward a most probable state (i.e. a graded channel [Leopold et al., 1964; Smith et al., 2000]). These theories have been widely accepted and thus, future fluvial process magnitude studies may rely on them also.

Fluvial erosion has already been defined as episodic by R.E. Horton (1875–1945), who further stated that changing circumstances induce rapid adjustments, in addition to progressive change (Kennedy, 1992). These episodic and progressive fluvial processes have been studied in both flume (i.e. physical models) and natural environments. In particular, both bed form evolution and particle movement studies have been commonly studied in the laboratory flume environment (e.g. Gilbert & Murphy, 1914; Bagnold, 1966, 1980; van Rijn, 1984a and b; van den Berg, 1987; Kuhnle, 1993; Roberts et al., 1998; Houwing, 1999; Ha and Maa, 2009; Ghoshal et al., 2010). Sediment transport laboratory flume studies were initiated in Germany by Engels (1854–1945) and in the USA by Gilbert (1843–1918) (van Rijn, 1993; Novak et al., 2001). One problem with laboratory flume studies is that scaling and artificial flow conditions can limit their applicability to the real world (Carling, 2000; Bridge, 2008). According to Bagnold (1977), the transport rates measured in small laboratory flumes cannot be applied to flows of greater depth without considerable modifications. Therefore, field studies are also needed.

At the end of the 1990's, there has been a trend more towards understanding and explaining river changes rather than their description (Lane & Richards, 1997). Field observations, such as river bathymetry, discharge, flow, bed sediment, bedload and suspended load measurements are needed to quantify a natural river's response to changes and also for sediment transport analysis (Helley & Smith, 1971; Ashworth & Ferguson, 1986; Lane et al., 1996; Carling et al., 2000; Kondolf et al., 2003; Duan & Scott, 2007). Nevertheless, it is difficult to make such measurements during high discharges, over large areas and long time periods (Leopold, 1980; Bridge, 2008). Flow measurements have greatly evolved from the mechanical current meters of the 1800s to the latest Acoustic Doppler Velocimeters (ADV) and Acoustic Doppler Current Profilers (ADCP), which are based on reflecting the sent acoustic signal from the particles moving in the water column (Whiting, 2003). Both ADV and ADCP have become popular due to their ease of operation in large rivers (Whiting, 2003;

García et al., 2005; Riley & Rhoads, online). At present, ADCP can also be applied to determine the bed load, based on the bias in bottom tracking due to bed movement (Rennie et al., 2002).

The correct representation of river bathymetry (i.e. geometry or topography) is important for both hydro- and morphodynamic studies. Point or cross-sectional echo-sounding data have been commonly applied in producing river bathymetry, particularly for multidimensional hydro- and morphodynamic simulations (Merwade, 2009). The developments in topographic data acquisition technologies (swathe sonar, airborne remote sensing, digital photogrammetry, differential GPS, laser profiling: LiDAR) enable faster and more detailed field measurements and geometry creation for fluvial studies, even in demanding large river reaches (Sirniö, 2004; Hicks et al., 2006; Hunter et al., 2007; Rumsby et al., 2008; Alho et al., 2009; Hohenthal et al., 2011; Flener et al., 2012).

2.1.2 Factors influencing channel hydro- and morphodynamics

River channels can be classified into three different categories: channels having either rigid (non-erodible), flexible (vegetated) or loose (erodible) boundaries (Knight, 1989). The rate of doing work has been established as the available power minus the unused power, or as the available power multiplied by the efficiency, i.e. the ratio of work done to the energy expended (Gilbert & Murphy, 1914; Bagnold, 1966). There are three forces concerned in traction: the force of the current, gravity and the bed resistance. Particles move when the fluid force, which is strongly related to near bed velocities, is larger than the resistance related to the weight, size, submergence and density of the bed material, channel slope and bed forms developed in the unstable loose boundary channels (Leopold, 1980; Komar, 1988; Copeland, 1990; Bathurst, 2002; van Rijn, 2007). After exceeding the critical threshold of motion, bed material moves by either rolling or sliding (Van Rijn, 1993). When the bed shear velocity increases and if the particle fall velocity is exceeded, finer sediments are lifted into suspension (Gilbert & Murphy, 1914; Komar, 1988; van Rijn, 1993). Although some views have recently emerged, which suggest that instead of excess shear, turbulent sweeps and ejections may be the primary triggering mechanisms of sediment entrainment (Papanicolaou et al., 2008), the widely studied threshold concept was applied in the present study.

The capacity of a stream has been defined as the maximum load it can carry and sediment is deposited when this capacity is exceeded (Gilbert & Murphy, 1914). Moreover, in coastal river reaches, it is well known that the backwater effect of the sea slows velocities and causes sedimentation, but erosion may also increase during high discharges when a draw-down of the fluvial water surface near the river mouth occurs due to the greater flow depth than water depth of the river mouth (Lamb et al., 2012). In particular, the backwater effect has been observed to occur during discharges smaller than those with an approximate two-year recurrence interval. The flow or stream competence may be applied as a measure of the ability of the current to move sediment (fluvial work) and is defined by the largest particles transported under a given set of flow conditions (Baker, 1977; Komar, 1989).

Alhtough the erosion rate of small particles (such as cohesive sediment) is known to decrease rapidly (i.e. the critical threshold for movement increases) as the bulk density increases, the erosion rate for the larger particles may be considered independent of bulk density (Roberts et al., 1998; Lick et al., 2004). The initiation of movement of particles smaller than 0.062 mm is affected by the binding forces between the particles: such cohesive properties become dominant when more than 5–10 % of the material is clay-dominated (<0.008 mm) (Krone, 1992; van Rijn, 2007). It is advisable to apply cohesive-type transport equations when particles smaller than 0.062 mm (silt) comprise more than 15 % of the mixture by weight (Mitchener & Torfs, 1996).

Flow parameters such as velocity, shear stress or stream power are commonly used to determine the critical bed load movement threshold (Hjulström, 1935; Chow, 1959; Bagnold 1977; Williams, 1983; Baker & Costa, 1987; Baker & Ritter, 1975; Knighton, 1984; Bagnold, 1986; Lick et al., 2004; van Rijn, 2007). For example, a critical shear stress or stream power is the threshold at which a small, but accurately measurable rate of erosion or particle movement occurs (Bull, 1979; Roberts et al., 1998). These thresholds have been defined for different grain sizes in different types of riverine environments. Instead of one certain critical threshold for a given particle grain size, zones of potential movement have been found for velocity, bed shear stress and stream power (Hjulström, 1935; Williams, 1983). Velocity is the oldest parameter for determining the threshold for the initiation of bed load movement (Hjulstrom 1935; Baker & Ritter 1975; Knighton 1984; Petit et al., 2005). However, Vanoni et al. (1966) stated that without specifying depth, velocity alone is insufficient for the initiation of particle movement. At present, shear stress applications have become common in geomorphology (Petit et al., 2005). Shear stress ($\tau_{bed,}$ N/m^2), here defined as bed shear stress, has been calculated in many fluvial studies for certain flow conditions as follows (Laursen 1958; Chow, 1959; Baker & Costa, 1987; Baker & Pickup, 1987; Rhoads, 1987; Magilligan, 1992; Ferguson, 2005; Aggett & Wilson, 2009; Parker et al., 2011):

$$\tau_{bed} = \gamma DS \qquad (1)$$

where γ is the unit weight of water (9800 N/m^3), which is calculated from ρ = fluid density (kg/m^3) and g = gravitational acceleration (m/s^2). Both the dynamic/absolute viscosity and density of water and therefore, the unit weight of water, are dependent on water temperature (Khorram & Ergil, 2010). Therefore, also the sediment transport rate has been shown to be effected by water temperature (Héquette and Tremblay, 2009). The slope (S) may be defined as either the energy slope (S_e, m/m) or may be approximated by the water surface slope (S_w, m/m) (Barker et al., 2009). It is possible to apply depth (D, m) instead of the hydraulic radius (R, m) if the width to depth ratio exceeds 10–20 (Chow, 1959; Rhoads, 1987; Magilligan, 1992). Further, the boundary shear stress can be determined if the equation is defined with the hydraulic radius (Chow, 1959; Baker & Costa, 1987; Baker & Ritter, 1975; Rhoads, 1987; Petit, 1990; Magilligan, 1992; Wilkinson et al., 2004; Kale & Hire, 2007; Aggett & Wilson, 2009):

$$\tau_{bo} = \gamma RS \tag{2}$$

In particular, shear stress has been applied for calculating the cohesive sediment movement (Ha & Maa, 2009; Houwing, 1999; Lick et al., 2004; Mitchener & Torfs, 1996; Roberts et al., 1998; Vanoni, 1966; Van Rijn, 2007; Winterwerp, 2007), since linking critical specific stream power to the material size has been found problematic (Petit et al., 2005). The unit stream power (W/m^2) is the amount of power expended by the flowing water per unit area of the channel bed (Costa, 1983). The experimental sediment transport rates were plotted against stream power (Eq. 3: Q = discharge m^3/s) for the first time by Bagnold (1966):

$$\Omega = \rho gQS \tag{3}$$

Stream power per unit channel length has also been referred to either as cross-sectional or total stream power (Barker et al., 2009) and has since been used widely (Baker & Costa, 1987; Phillips & Slattery 2007; Barker et al., 2009). Although both the flow strength and bedload transport vary within curved channels, this is ignored in approaches where the mean shear stress, competence or capacity are applied (Ferguson, 1987). An equally widely applied form of the equation exists (Eq. 4: Bagnold, 1966), where local variations of depth (D) and velocity (v) have been introduced instead of discharge (Bull, 1979; Costa, 1983; Rhoads, 1987; Magilligan, 1992; Lewin & Brewer, 2001):

$$\omega = \rho gDSv \tag{4}$$

Sediment transport algorithms commonly include velocity, shear stress or stream power components and their thresholds for incipient motion (Mosselman, 2005). While the first bed load type formulae based on depth and slope were presented by Du Boys in 1879, the first empirical formula was developed by Meyer-Peter & Müller in 1948 (van Rijn, 1993). Other workers, for example, Einstein (1950), Laursen (1958), Krone (1962), Partheniades (1965), Bagnold (1966, 1980), van Rijn (1984, 1989, 1993, 2007) and Copeland (1990) have also developed equations for non-cohesive or cohesive sediment transport. These algorithms have been validated in certain environmental and sediment transport conditions and therefore, are valid where the certain conditions hold (Bathurst et al., 1987; Neumeier et al., 2008; Brunner, 2010). Most of the bed and suspended load transport algorithms have been created for non-cohesive sediments composed of sand-sized particles. Despite multiple re-testing and re-interpretation of the transport equations over the years (Miller et al., 1977; Leopold, 1980; Martin & Church, 2000; Ferguson, 2005; Wong & Parker, 2006; Winterwerp, 2007; Parker et al., 2011), further trials of different algorithms in different environments are still needed.

Two of the most commonly applied sediment transport equations, which include the threshold for incipient motion, in morphodynamic studies are the formulae of

Meyer-Peter & Müller (MPM) (Meyer-Peter & Müller, 1948; van Rijn, 1986; Martín-Vide et al., 2010) and van Rijn (VR) (van Rijn, 1984a–c; van den Berg, 1987; van Rijn, 1989; van Rijn, 2007; Ghoshal et al., 2010). The MPM bed load formula, which has been considered appropriate for example for wide channels (Chanson, 1999), can be defined as the original and widely used 1948 form of the relationship:

$$\rho_w\left(\frac{K_s}{K_r}\right)^{\frac{3}{2}} \times h \times S_b = (0.047 \times \rho_w (s-1) \times g \times D_m) + \rho_w (s-1)^{\frac{2}{3}} \times g^{\frac{2}{3}} \times q_b^{\frac{2}{3}} \quad (5)$$

where:
q_b = sediment discharge rate

s = relative density = $\frac{\rho_s}{\rho_w}$
ρ_s = sediment density
ρ_w = water density
h = water depth
S_b = bed slope
D_m = effective grain diameter, such as D_{50}
g = acceleration due to gravity
K_s = Strickler roughness coefficient for the bed material in question
K_r = Strickler roughness coefficient related to particle friction only (i.e. skin friction)

Van Rijn's relationships, which have been developed for sand and gravel particles, include both bed and suspended load transport (van Rijn, 1984a–c). The bed load relationship has been defined in literature as follows (van Rijn 1989):

$$q_{b,c} = \frac{0.25 \times u'_* \times D_{50} \times T^{\frac{3}{2}}}{D_*^{0.3}} \quad (6)$$

where:

u'_* = grain related bed shear velocity = $\frac{\sqrt{g} \times \bar{u}}{C'}$

C' = grain related Chezy coefficient $= 18\log\left(\frac{12h}{3D_{90}}\right)$

D_* = dimensionless particle parameter = $D_{50} \times \left[\frac{((s-1)g)}{v_k^{\ 2}}\right]^{\frac{1}{3}}$

T = transport stage parameter = $\frac{\tau'_b - \tau_{b,cr}}{\tau_{b,cr}}$

$\tau_{b,\,cr}$ = critical shear stress in accordance as according to Shields (1936)

$$\tau'_b = \frac{1}{8}\rho_w f'_c \bar{u}^2$$

f'_c = grain related Darcy Weisbach friction factor = $\dfrac{0.24}{\log\left(\dfrac{12h}{3 \times D_{90}}\right)^2}$
g = acceleration due to gravity
$\bar{u}$ = depth averaged velocity
h = water depth
ρ_w = water density
v_k = kinematic viscosity

Although small-scale bed forms, such as ripples are known to respond rapidly to the changing discharges, large bed forms (macro-scale), such as dunes, respond slower and a phase lag between the new flow conditions and the establishment of the new dune conditions may occur (van Rijn, 1993). Channels with sand beds and cohesive banks may be particularly susceptible to modest increases in stream power (Bledsoe & Watson, 2001). When velocities are moderately (10–20 %) higher than the critical velocity for initiating motion and the particle size is smaller than 500 µm, small ripples are generated at the bed surface (van Rijn, 1993).

Peak discharges, flow velocities, slope and width-depth ratios, specific stream power and shear stresses are normally greater in braided rivers than in straight or meandering channels (Parker, 1976; Schumm, 1985; Ferguson, 1987; Kleinhans & van den Berg, 2011). The development of braiding is favoured if the bed load transport is high relative to other transport modes and the bank material is erodible, so that a wide, shallow channel may develop (Carson, 1984; Summerfield, 1991; Murray & Paola, 1994). When discharges decline and coarse material is deposited, a nucleus for a sand bar forms. The bars develop when further amalgamation causes them to grow (Summerfield, 1991; Ashworth et al., 2000; Sambrook Smith, 2006). More recently, it has been stated that in fact flow instability and a high width to depth ratio may also cause a single-thread velocity field to break into multiple flow threads that deposit material between them and result in braiding (Richardson & Thorne, 2001). Further work for establishing a link between these processes operating in braided rivers and the changing morphology through successive flood events has been suggested (Sambrook Smith, 2008).

2.1.3 Different magnitude discharge events as channel modifiers

For several decades, the knowledge of the present and future magnitudes, as well as the probable frequency of flood occurrence has been considered important due to the geomorphic effects of discharges (Dalrymple, 1960; Leopold et al., 1964). Such information is also necessary for selecting the proper design and location of structures, such as dams and bridges, as well as flood inundation studies. The recurrence interval or flood frequency has been defined as the average interval of time within which a flood of a given magnitude will be equalled or exceeded once and is also a statement of probability (Leopold et al., 1964; Ward, 1978). For example, a 100-year discharge is a discharge with a recurrence interval of a century, i.e. it has a 1 % chance of recurring in any single year (Dalrymple, 1960; Leopold et al., 1964).

Fisher and Tippett (1928) developed frequency distributions of maximum values, which were subsequently applied to floods by Gumbel in 1945 (Dalrymple, 1960). Most methods used in the probability analysis of flood flows are based on the annual flood series, i.e. the highest flood in each year of the data period (Ward, 1978).

The changes caused by floods are effected not only by the resistance of the river channel but also the magnitude and frequency of the input forces (Baker, 1988). Thus, the effectiveness of flood erosion depends on exceeding a critical resistance threshold, such as those defined for shear stress and stream power. For example, the minimum thresholds of bed shear stress and unit stream power for catastrophic flooding have been defined as 100 N/m^2 and 300 W/m^2, respectively (Magilligan, 1992). The maximum discharges may create new boundary conditions for sediment transport, in addition to the changes caused to the slope and river bed morphology and particularly to erosion (Ward, 1978; Osterkamp & Costa, 1987; Gintz et al., 1996).

At present, there is ongoing discussion concerning whether in certain river environments it is the high discharges that are the most important river channel modifiers or if the greatest changes are caused by the longer lasting low discharge periods. Although high discharges may increase erosion, they occur rather infrequently. For example, Baker (1977) has shown that local characteristics affect the geomorphic responses to events. In addition, the great impacts of infrequent high floods, such as 100-year discharges, have recently been found to be highly stream specific (Fuller, 2008). The variability of geomorphic impacts and whether a channel shows a catastrophic response can be explained with hydrological, hydraulic and geomorphic variables, such as drainage basin size, basin morphology, drainage density and the channel resistance (Baker, 1977; Fuller, 2008). It has been noted that in determining the geomorphological effects on channels, both the sediment availability and sequence of events may be as important as the flood magnitude (Carling & Beven, 1989). Moreover, the conditions during the flood event in questions, e.g. if there are already erosion scars from preceding years, influence the actual sediment transport (Bogen, 2009).

Another view that has been proposed is that while extremely high but infrequent flows are effective in geomorphic work (erosion and transport), their total effect on channel morphology is less than smaller but frequently occurring discharges (Leopold et al., 1964). A channel-forming discharge is one which if maintained indefinitely, would produce the same channel geometry as the natural long term hydrograph (Copeland et al., 2000). Such a flow has been defined by means of either a bank-full (maximum discharge without overflowing) or effective discharge, which transports the greatest proportion of the annual sediment load over a period of years. In many rivers, the recurrence interval of the bank-full stage has been found to be approximately 1–2 years, or 1.5 on average (Leopold et al., 1964). However, in other rivers the discharge corresponding to the 1.5-year recurrence interval has not represented the bank-full discharge (Williams, 1978). It is interesting to note that the effective discharges of return periods 1.2–1.45 years have been found to be those discharges at which, over a period of time, the most bed-load is transported (Pickup, 1976). Further, later studies have indicated that a flood does not need to have extreme values of precipitation, unit discharge or unit stream power to cause catastrophic change on the valley floor, while floods with these extreme values have been shown to cause only minor changes (Miller, 1990; Magilligan et al. 1998; Arnaud-Fassetta, 2003). Although a high discharge event, e.g. a 40-year flood, may rework the entire braid-plain, both the scour and deposition may be similar to those associated with lower magnitude annual floods for example in braided sandy rivers (Sambrook Smith et al., 2010).

In any case, it has been found necessary to further improve the understanding of sediment delivery during extreme situations, in order to predict the potential effects of climate and human impacts on sediment fluxes (Bogen, 2009). Due to the fact that different rivers respond differently to different magnitude discharges, further case studies on climate change impacts on discharges and their morphodynamic effects are needed from different riverine environments. Since the relationship between the channel modifications and flow characteristics has been established, it is possible to investigate the temporal and spatial variations of the morphological impacts of different magnitude discharges by applying calculations of velocity, shear stress and stream power.

2.1.4 The observed impacts of climate on fluvial processes

Global hydrological change has been said to result from the combined effects of climate change, land use, water transfer and river engineering (Meybeck et al., 2003). In addition, the climate is affected by changes in the atmospheric abundance of greenhouse gases and aerosols, in solar radiation and in land surface properties (IPCC, 2007). Moreover, variations in climate, flood frequency and land-use pattern that have already occurred in the past have been recorded in the terrace and floodplain sediments or in the stratigraphy of large rivers (Carling & Beven, 1989; Benito et al., 2010). One of the effects of climate has thus been the changes in discharges. The alluvial records of paleofloods show that natural floods resulting from excessive rainfall and snowmelt have been highly sensitive to even modest changes of climate (Knox, 2000).

During the 20th century, the frequency of 100-year floods has been found to increase in basins larger than 100 000 km^2, and an increasing flood risk is forecast to continue in the future (Milly et al., 2002). In Nordic regions, case studies in smaller watersheds have been performed to detect the past climate impacts on discharges. According to Bogen (2009), the seasonal distribution of run-off in several parts of Norway has already changed in recent times and higher winter temperatures have produced an increase in winter floods, as well as the frequency of flood events. While the analysis of long term discharge records of 25 rivers or lake outlets in Finland (1912–2004) has shown seasonal stream flow changes, no change in mean annual flow in general has been found (Korhonen & Kuusisto, 2010). In particular, an increase in the winter and spring mean monthly discharges have mostly been observed, with spring peak discharges occurring earlier.

Even though discharge changes exert influence on the fluid forces, sediment transport and channel morphology, there are still relatively few studies of such impacts. An increase in the average boundary shear stress and specific stream power, together with increasing water depths and discharges have been reported in the Rhône delta within a 100-year period from 1895 to 1995 (Arnauld-Fassetta, 2003). Further, Bull (1979) has stated that the critical power threshold is sensitive to changes in climate, base level and the impact of humans. In turn, such threshold changes can cause aggradation or degradation. Recently, Amsler et al. (2005) also found a link between climate fluctuation, hydrology and channel morphology, since periods of different effective discharges of the 20th century correlated fairly well with climate fluctuations in the Paraná River: morphologic parameters, such as mean width, thalweg sinuosity, braiding index and aspect ratio increased during high discharges. Thus, the climate can also be expected to impact on fluvial process magnitudes in the future.

It must be kept in mind that in addition to climatic changes, human influences, such as on land use and channelisation, can sometimes cause major changes to the fluvial processes and morphodynamics (Trimble, 2003; Thorndycraft & Benito, 2006; Benito et al., 2010). For example, hydro-climatic changes and human disturbances in the Rhône catchment, such as sediment dredging and the building of dams, have been noticed to cause sediment yield reductions and increased stream power and shear stress, accompanied by geomorphological adjustment and channel incision (Arnaud-Fassetta, 2003).

2.2 A review of simulation approaches

2.2.1 Hydrological simulation approach

Hydrological models are applicable for gaining more information from hydrology and run-off processes in general, but also from past, future and rarely occurring hydrological events (Freeze & Harlan, 1969; Raudkivi, 1979). Models can be divided into high- (i.e. spatially distributed) and a low-resolution (i.e., spatially lumped, conceptual) models (Refsgaard, 1997; Carpenter & Georgakakos, 2006). In the lumped models, the catchment is considered as a single unit characterised by a few tens of parameters and variables, while the distributed models are structured as a network of thousands of grid points, each of which have several parameters and variables (Refsgaard, 1997). Distributed models have received criticism, since there are many parameter values that may be modified during calibration (Beven, 1989; Refsgaards, 1997).

A successful hydrological model has been stated to require the proper accounting for soil moisture (Bergström et al., 2001). The HBV model (Hydrologiska Byråns Vattenbalansavdelning; Bergström, 1976), whose development began in 1972 and which was first published in 1973 by Bergström and Forsman, was one of the first hydrological models to adopt soil moisture variability parameter. The HBV model started as a very simple lumped hydrological model and has become more distributed with time (Bergström et al., 2001). For example, the Watershed Simulation and Forecasting System (Vehviläinen et al., 2005), a HBV-type conceptual model of the Finnish Environment Institute (FEI), is used in Finland for operational hydrological and flood forecasting, flood warnings and climate change research (e.g. Vehviläinen

& Huttunen, 1997).

Together with increasing awareness of climate change, the hydrological regional or national scale simulations of climate change impacts on discharges have increased, particularly within the last 10–20 years (e.g. Bergström et al., 2001; Arnell, 2003; Andréasson et al., 2004; Lehner et al. 2006; Graham et al., 2007; Beldring et al., 2008; Dankers & Feyen, 2008; Steele-Dunne et al., 2008). When climate change impact is applied in hydrological simulations, the emission scenario is used to run a global climate model (GCM) simulation, which is possibly down-scaled (Regional climate model: RCM), bias-corrected, and then applied to the hydrological model (Seguí et al., 2010). The 1990–2100 emission estimates for important radiative gases have been generated in the IPCC Special Report on Emission Scenarios (SRES) (IPCC, 2000). These have been grouped into six scenario groups (A2, A1B, A1FI, A1T, B2, B1), of which the A2 and B1 scenarios, respectively, have high and low emission levels. By applying these emission scenarios together with the climate models, the future climate scenarios and thus, temperature and precipitation estimates, can be obtained for the hydrological simulations of future discharges.

The data produced by global climate models (GCM) may be unsuitable to feed directly into the hydrological models, partly due to the mismatch between the climate model output, which has a coarse spatial resolution, and the spatial scale of hydrological models. To counteract this, down-scaling procedures have been developed (Wood et al., 2004; Leander & Buishand, 2007). In addition, different bias correction methods of direct RCM data have recently been generated, as well as other transfer methods have been developed for transferring the climate change from the climate models to the hydrological models (Wood et al., 2004; Lenderink et al., 2007). Despite the limitations in hydrological models, e.g. related to future snow and evapotranspiration modelling, it has been stated that the applied climate scenarios themselves are the main source of uncertainty in climate change impact simulations (Bergström et al., 2001). Therefore, further fluvial studies with multiple different climate and thus, hydrological scenarios are also needed.

2.2.2 Regional and national scale future discharge change forecasts

While future warming has been forecast to be particularly high in northern latitudes, future precipitation increases are also very likely (IPCC, 2007). Due to the complex nature of climate change, studies of climatic change impacts have been acknowledged to be among the most complicated and uncertain environmental assessments (Kuikka & Varis, 1997). Although simulations of future discharge changes on a regional scale, such as a continent, have been performed, some inconsistencies occur. An integrated continental analysis of the possible impacts of global climate and water use change on future flood and drought frequencies for Europe was first presented by Lehner et al. (2006). Based on their study, the most prone regions to a future rise in flood frequencies are forecast to be in northern and north eastern Europe (including Finland). Graham et al. (2007) also indicated an overall increase in river flow, earlier spring peak flows and an increase in the potential for hydroelectric power. Conversely, Dankers & Feyen (2008) found a considerable decrease in future flood hazards in north eastern Europe, where warmer winters and shorter snow seasons reduce the magnitude of the spring snow melt peak. More locally, although a great variation in the future total mean annual river flow change to the Baltic Sea is shown between scenarios, a generally increasing future river flow trend has been forecast in the north of the Baltic Basin (Graham, 2004).

Studies of climate change impacts on discharges have been performed on a national scale in both central Europe and in the Nordic region (Arnell & Reynard, 1996; Steele-Dunne et al., 2008; Prudhomme et al., 2003; Dankers et al., 2007; Lenderink et al., 2007). Although, it has been forecast in Norway that snow melt floods will occur earlier in the year and that their occurrence will become more seldom due to earlier snow melt and reduced snow storage, the increased autumn and winter rainfall will increase stream flows, while the summer discharges will decrease (Beldring et al., 2006; Beldring et al., 2008). Spatial variations in future spring discharges have been simulated both in Norway and Sweden according to whether the basin is in the northern or southern part of the country (Andréasson et al., 2004; Beldring et al., 2006). In southern Sweden decreasing spring floods, summer run-off and annual run-off have been projected, while annual run-off has been forecast to increase in northern Sweden (Andréasson et al., 2004). Bergström et al. (2001) also forecast a general decrease in flood risk in southern Sweden and less dominant spring floods, together with possible shifts of extreme run-off from spring to other seasons in northern rivers. In Finland, it has been predicted that future precipitation and temperature increases will change the seasonal distribution of run-off: winter precipitation will increase, spring floods will diminish or vanish and be replaced by winter floods (Vehviläinen & Huttunen, 1997). Winter floods have been forecast to increase the most in the central lakes. Recently, support for an increasing mean winter run-off and an earlier spring snow melt spring discharge peak has also been gained, based on different climate scenarios (Veijalainen et al., 2012).

Therefore, due to the above mentioned inconsistencies of regional scale studies and great spatial variability in future predictions on a national scale, the need for more detailed studies are obvious. Studies, such as Verhaar et al. (2010), point out the importance of investigating several rivers from different regions, using several climate models to determine trends in climate change impacts. It is also important to acknowledge that when natural variability is added to the human-induced climate change, the ranges of possible future stream flows and uncertainties may increase substantially, although on some occasions the natural variability may counteract the climate change also (Arnell, 2003).

2.2.3 Hydro- and morphodynamic simulation approaches

It has been said that the most desirable approach to the numerical modelling of open channel flows and hydrodynamics is through the solution of the fundamental equations of fluid and sediment motion (conservation of mass and momentum) (Bridge, 2008). The spatial numerical solution techniques of these equations, i.e. Navier-Stokes equations, and their computer implementation developments since the 1950's and 1960's, enable the calculation of fluid motion in geometrically complex river channels with spatially and temporally varying boundary conditions (Chanson, 1999; Hardy et al., 1999; Bates et al., 2005; Sambrook Smith et al. 2006). The governing equations can be solved when a model's boundary conditions, commonly discharge and water level conditions, are specified (Writgh, 2005; Lane et al., 2008).

Although early applications of CFD were mainly focussed on steady state solutions, unsteady simulations are now common (Wright, 2005). An unsteady flow, which is opposite to steady flow concept, is defined as a flow that changes with time. In a non-uniform flow, the river cross-sections and water levels do not remain unchanged throughout the river reach, as occurs in uniform flow (Olsen, 2007). Fluid particles move horizontally and different water layers may also become mixed if the flow is turbulent. In a sub-critical flow situation, where the flow velocity is smaller than the speed of the disturbance wave, the model calculations should be started from the downstream side and continued upstream, due to the effects of surface wave disturbance (Wright, 2005). The main established parameters applied in hydrodynamic simulation studies include the channel geometry (grid, mesh or cross-section form), the properties of the flowing fluid (density, viscosity), the flow characteristics (velocity, flow depth, water level), as well as estimates of resistance and roughness (Chanson, 1999; Hunter et al., 2007).

Hydrodynamic simulations can be performed either in one- (1-D: Wilkinson et al., 2004; Aggett & Wilson, 2009), two- (2-D: Hunter et al., 2008) or three-dimensions (3-D: Booker et al., 2001; Rodriguez et al., 2004). 1-D models calculate flow only into one direction, i.e. downstream, and the calculations and results are averaged cross-sectionally (Ferguson, 2003; Nelson et al., 2003; Bates et al., 2005; Carling et al., 2010). The advantage of 1-D models is their ability to simulate changes over larger areas and longer time spans (reaching also into the future) compared to multidimensional models, which require more computing power (Formann et al., 2007; Verhaar et al., 2010). While 2-D models solve shallow water equations for water level and depth-averaged velocities in two spatial directions, these results are vertically averaged (Nelson et al., 2003; Pender & Néelz, 2006). 3-D models can also vertically simulate the secondary flows and momentum flux variations (Nelson et al., 2003). Therefore, if it is required to simulate flow, sediment transport, erosion and deposition at certain locations, multidimensional 2-D or 3-D models can be better applied in flood plain environments, confluences and meandering river reaches (Nelson et al., 2003; Pender & Néelz, 2006; Carling et al., 2010). According to some authors, 2- or 3-D models should be used in braided rivers (Nicholas, 2000; Bertoldi et al., 2009; Brunner, 2010). The numerical solutions for the equations of 2- and 3-D models first became available in 1960's (Hunter et al., 2007). Nevertheless, both 1- or 2-D models were still most widely applied at the beginning of the 21st century (Lesser et al., 2004). Thus, the choice of the numerical model depends on the complexity of the processes to be analysed, the quality of the field data, and the computational limitations (Formann et al., 2007).

The continuity equations (i.e. the partial differential equations of the conservation of mass [Eq.7] and momentum [Eq. 8]) can be written in 1-D (HEC-RAS model: Brunner, 2010) as follows:

$$\frac{\partial A}{\partial t}+\frac{\partial S_s}{\partial t}+\frac{\partial Q}{\partial x}-q_1=0 \qquad (7)$$

where:
x = distance along the channel
t = time
Q = flow
A = cross-sectional area
S_s = storage from non conveying portions of cross section
q_1 = lateral inflow per unit distance

$$\frac{\partial Q}{\partial t}+\frac{\partial (vQ)}{\partial x}+gA\left(\frac{\partial \zeta}{\partial x}+S_f\right)=0 \qquad (8)$$

where:
g = gravitational acceleration
S_f= friction slope
v = velocity
ζ = water surface elevation

In addition, in the 2-D model (TUFLOW model: BMT WBM, 2008) the conservation of mass (Eq. 9) and momentum (Eq. 10, X momentum; Eq. 11, Y momentum) equations are depth-averaged and are definable as follows:

$$\frac{\partial \zeta}{\partial t}+\frac{\partial (hu)}{\partial x}+\frac{\partial (hv)}{\partial y}=0 \qquad (9)$$

$$\frac{\partial u}{\partial t}+u\frac{\partial u}{\partial x}+v\frac{\partial u}{\partial y}-c_f v+g\frac{\partial \zeta}{\partial x}+g\,u\left(\frac{n^2}{h^{4/3}}\right)\sqrt{u^2+v^2}-\mu\left(\frac{\partial^2 u}{\partial x^2}+\frac{\partial^2 u}{\partial y^2}\right)+\frac{1}{\rho_w}\frac{\partial p_a}{\partial x}=F_x \qquad (10)$$

$$\frac{\partial v}{\partial t}+u\frac{\partial v}{\partial x}+v\frac{\partial v}{\partial y}+c_f u+g\frac{\partial \zeta}{\partial y}+g\,v\left(\frac{n^2}{h^{4/3}}\right)\sqrt{u^2+v^2}-\mu\left(\frac{\partial^2 v}{\partial x^2}+\frac{\partial^2 v}{\partial y^2}\right)+\frac{1}{\rho_w}\frac{\partial p_a}{\partial y}=F_y \qquad (11)$$

where:
ζ = water surface elevation
u and v = depth averaged velocity components in X and Y directions
h = water depth
g = gravitational acceleration
t = time
x and y = distance in X and Y directions
Δx and Δy = cell dimensions in X and Y directions
c_f= Coriolis force coefficient

n = Manning's n
μ = horizontal diffusion of momentum coefficient
p_a = atmospheric pressure
ρ_w = density of water
F_x and F_y = sum of the components of external forces (e.g. wind) in X and Y directions

Since their first appearance, numerical hydrodynamic models have been especially applied for studying spatial and temporal variations in flow characteristics (Booker et al., 2001; Nicholas, 2003; Pasternack et al., 2006; Phillips & Slattery, 2007), present or observed flood inundation areas (Horrit & Bates 2002; Néelz et al., 2006; Pappenberger et al., 2006; Hunter et al. 2007; Koivumäki et al., 2010), flood plain processes (Hardy et al., 1999; Wilson et al., 2006) and mega floods (Carling et al., 2010). In addition, using numerical simulations, more knowledge has been gained from different types of river reaches, such as meandering (Duan, et al., 2002; Minh Duc et al., 2004; Duan & Nanda, 2006; Rüther & Olsen, 2007), braided (Fassnacht, 1997; Hicks et al., 2006; Nafziger et al., 2009) or confluent/bifurcating channels (Dargahi, 2004; Duan & Nanda, 2006). The continuous refinement and development of numerical models increases their computing capabilities and enables their applications in larger areas and longer time spans with increasing detail. Changes in the future flow characteristics and thus, the erosion and sedimentation potential have been detected in only a few studies and there exists a need for future fluvial studies with numerical models (Verhaar et al., 2011). In addition, more detailed information for a better understanding of the combined effects of present and future discharges and sea water levels on the inundation processes and fluid forces exerted on the river bed is also needed, not least because of the demands to incorporate them in future flood hazard or risk estimations (European commission, 2007; Lane et al., 2007). Numerical simulations make these detailed studies possible.

The evolution of river channels can be simulated using morphodynamic models, which include the traditional equations for the initiation of sediment particle motion (Meyer-Peter & Müller, 1948; van Rijn, 1984a–c), the conservation of sediment (e.g. the Exner equation: Benkhaldoun et al., 2011), as well as the bed update scheme (Callaghan et al., 2006). For example, the development of 1-D sedimentation models was begun in 1967 by the US Army Engineer District (Copeland, 1990).

Although the actual morphodynamic simulations of the river bed (including bed load transport and bed update) have been performed in different dimensions (e.g. Copeland, 1990; Simon & Darby, 1997; Rathburn & Wohl, 2003; Dargahi, 2004; Nagata et al., 2005; Bertin et al., 2009), in some studies only the suspended sediment transport is simulated, based on different sediment transport formulae or sediment rating equations (Fassnacht et al., 1997; Morehead et al., 2003; Duan & Nanda, 2006; Formann et al., 2007). Morphodynamic simulations have rarely been applied to long time periods and large spatial scales or where there are complicated temporal and spatial variations in geometry, water and sediment supply (Dargahi, 2004; Lesser et al., 2004; Bridge, 2008). Nevertheless, studies of the geomorphic response to sea level rise and climatic variability in the 21st century (Ganju & Schoellhamer, 2009), as well as the morphodynamic simulation tests to model a 100-year future period by Verhaar et al. (2008) give confidence to the reliable application of numerical models when solving climate change impacts on river morphodynamics.

In addition to numerical morphodynamic models, cellular models have also developed and their usage, especially in braided river studies, has increased (Murray & Paola, 1994; Doeschl-Wilson & Ashmore, 2005; Doeschl et al., 2006; Nicholas et al., 2006; Sambrook Smith et al., 2006; Coulthard et al., 2007; Hodge et al., 2007). Although it is possible to derive the equations from 1- and 2-D numerical models to cellular ones (Doeschl et al., 2004), the actual numerical morphodynamic models have more sophisticated rules and may contain significant empirical components (Nicholas et al., 2006).

All morphodynamic models need measured sediment properties and transportation amounts in addition to the other field measurements required by hydrodynamic models (Nicholas, 2003; Bridge, 2008; Hu et al., 2009). The horizontal and vertical bed material composition and variation have often been neglected in sediment transport and morphological models at the beginning of the 21st century (van Ledden, 2002). Moreover, the grain size sorting and temporal variation is only included in few models (Verhaar et al., 2011). The widely applied approach of using a single fixed grain size in sediment transport equations may not represent the various size fractions within sediment mixtures (Wu et al., 2004). Although the morphological modelling of cohesive sediment transport has been performed for different applications and river environments (Liu et al., 2002; Rathburn & Wohl, 2003; Lopes et al., 2006; Neumeier et al., 2008; Verhaar et al., 2008; Verhaar et al., 2010), the nature of cohesive sediments complicates their simulation (Krone, 1962; Partheniades, 1965; van Ledden, 2004, 2002; Verhaar et al., 2008). The choice of transport algorithm has been found to be important (Bertin et al., 2009). Since it has also been stated that sediment transport models may not be as universal as thought (Papanicolaou et al., 2008), researchers are encouraged to employ the applications of different models with a selection of algorithms in different types of river environments.

2.2.4 Forecasts of future sediment transport and morphodynamics

Although the impacts of changing discharge events in the future on sediment load, and in particular, bed load, have been simulated in only few studies (Verhaar et al., 2011), some examples from the Nordic and high latitude regions exist. For example, in Denmark it has been forecast that although suspended sediment transport will increase during the winter months, due to the increasing river discharge caused by the rise in precipitation, it will decrease during the summer and early autumn months (Thodsen et al., 2008). The mean annual suspended sediment transport is expected to increase less in an alluvial than in a non-alluvial river (Thodsen et al., 2008). In addition, an increasing average bed material delivery has been predicted in the Saint-Lawrence River, Canada (Verhaar et al., 2010). Similarly, according to Gordeev (2006), a large sediment flux increase (i.e. 30–122 %) is projected for six of the largest Arctic rivers in Russia (i.e. Yenisey, Lena, Ob, Pechora, Kolyma and Severnaya Dvina) by 2100.

Verhaar et al. (2010) noticed that the magnitude of simulated discharge and bed material delivery change depends on the choice of the global climate model. Much variation has also been shown in the future suspended sediment transport changes (Gomez et al., 2009). Due to this variation between regional, national and local scale studies, it is important to detect the effects of climate change on sediment transport and channel modifications using many different scenarios.

The climate change is suggested to have a joint impact with sedimentation on inundation areas and flood risk with the possibility that severe and combined effects of climate change and sediment delivery may occur (Lane et al., 2007). The aggradation of a river bed may potentially cause a loss of capacity and influence flooding (Gomez et al., 2009). According to Anisimov et al. (2008), in combination with other environmental changes, the climatic warming may also lead to a transformation of the river channel types. Due to the scarcity of simulations on future sediment transport changes and morphodynamics, there exists an urgent need to further investigate these issues (Verhaar et al., 2011).

3 Study areas

3.1 Climatic and hydrological characteristics

Since Finland is influenced by both the maritime climate from the Atlantic Ocean, as well as the continental climate from the Eurasia, its climate can be classified as cold, but without dry season (Df, Köppen-Geiger climate classification: Peel et al., 2007; Kersalo & Pirinen, 2009). While the most southern coast has warm summers (Dfb), they are cool in the rest of the country (Dfc). July is the warmest month and during the period 1971–2000, the average annual temperatures have varied from +5 to -2 ºC, with precipitation from 450 to 700 mm (Drebs et al., 2002; Kersalo & Pirinen, 2009). These average temperature and precipitation ranges have also been rather similar in the 1981–2010 period (Fig. 2). The highest and lowest mean annual temperatures occur in the south west and in the north east, respectively. The mean annual temperatures in Finland have increased by approximately 0.7 ºC in the period 1901–2000 (Jylhä et al., 2004). The observed linear temperature trend in spring and summer has been +1.4 and +0.7 ºC/100 years, respectively. There has only been a modest yet statistically insignificant warming in autumn, while winter temperatures have shown large variations. The highest precipitation has occurred during July-August, with the least during winter and spring (Kersalo & Pirinen, 2009). The average maximum snow water equivalents vary in southern and northern Finland from 80–140 mm to 140–200 mm, respectively, with an average snow thickness in excess of 80 cm in the latter (Fig. 2). In southern Finland, the spring run-off due to snow melt and rain has been 40–50 % of the yearly run-off, and as much as 50–60 % in northern Finland.

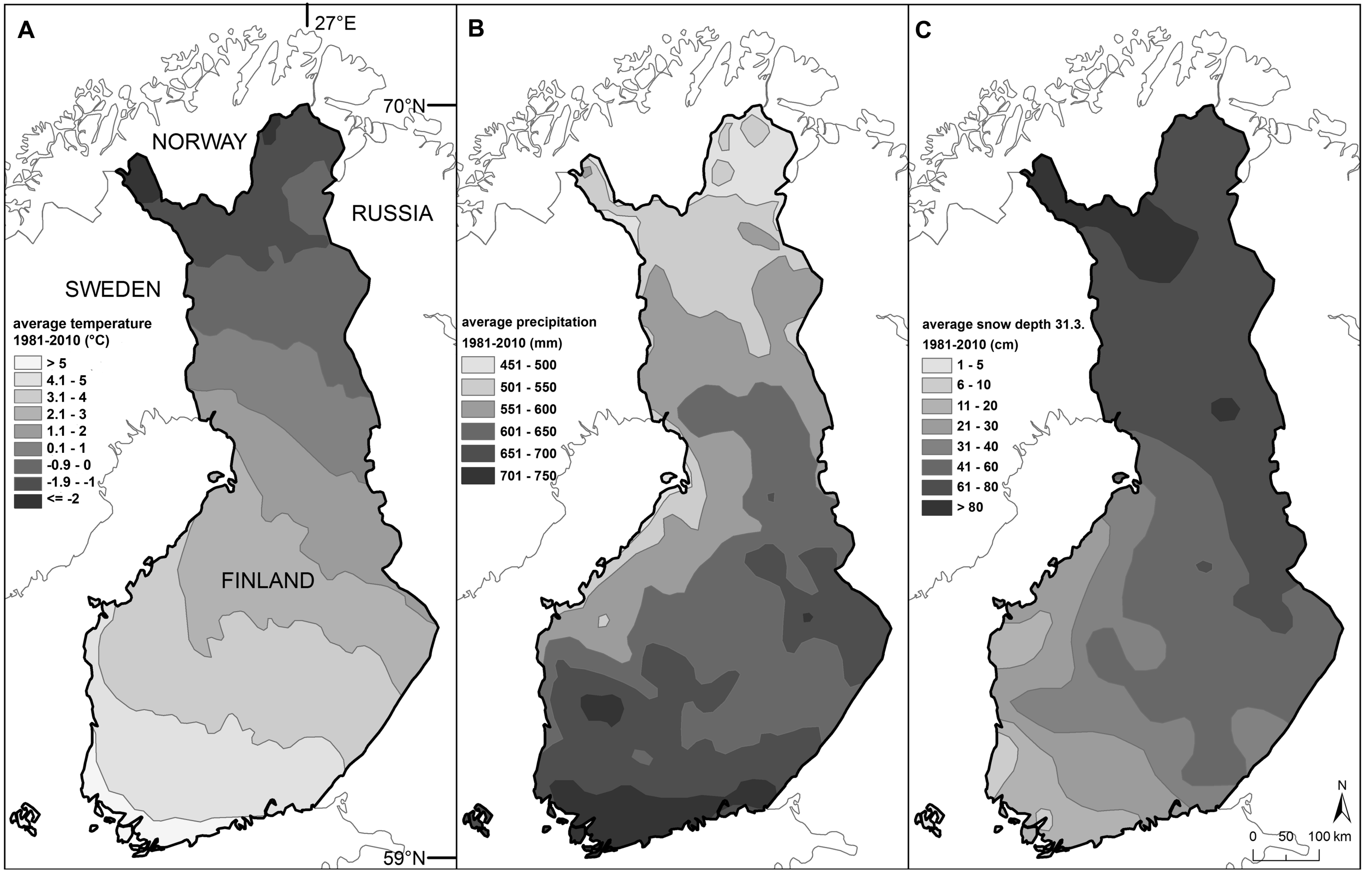

Fig. 2. A) The average temperatures in the period 1981–2010, B) the average precipitation for the period 1981–2010, as well as C) the average snow depth on the 31st March during the same period (follows the delineations by the Finnish Meteorological Institute, 2012). The snow cover is normally its thickest in March. Institute, 2012). The snow cover is normally its thickest in March.

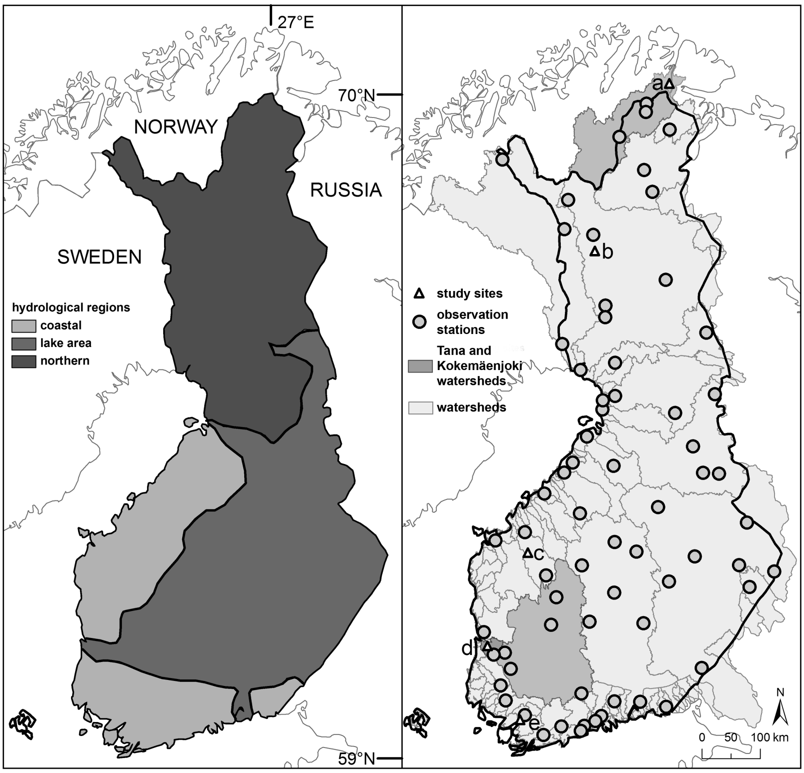

Fig. 3. A) The hydrological regions of Finland (follows the division by Korhonen & Kuusisto, 2010). B) The detailed hydrodynamic or morphodynamic simulation study sites of papers I–IV: a=Tana River (Seida settlement), b=Ounasjoki River (Kittilä settlement), c=Lapuanjoki River (Lapua settlement), d=Kokemäenjoki River (city of Pori), e=Uskelanjoki River (city of Salo). The 67 observation stations of paper I and the main watersheds of Finland are also shown. The Tana River and Kokemäenjoki River watersheds are presented as darker grey.

The 74 main watersheds in Finland can be divided into coastal, lake area and northern watersheds (Fig. 2) (Mustonen, 1986). In central Finland, lakes are especially numerous, with approximately 40 lakes per 100 km^2 and a local lake percentage of the area in excess of 35 %. In total, there are 188 000 lakes of at least 0.05 ha, i.e. 10% of the total area of Finland. The largest lakes, which exceed 1000 km^2, include Saimaa (lake area), Inari (northern) and Päijänne (lake area). The coastal region is characterised by small and medium sized rivers, while large and medium sized rivers also exist in northern Finland. The greatest discharge event of the hydrological year, particularly in the northern rivers, is the spring snow melt discharge. The 67 study sites of paper I represent many watershed types and sizes (Fig. 3). The selection criteria for these sites included the length and quality of the discharge observations, as well as their relative independence from each other. More detailed flood discharge analyses were performed at five sites representing different hydrological regions, while future flood inundation analysis was performed in four river reaches representing different watershed categories. Major flood damages have occurred in these four locations, which are also listed as flood prone areas by Timonen et al. (2003). The national flood hazard mapping with 20–1000 year floods has been carried out at these sites by both the Finnish Environment Institute and Regional Environment Centres.

The approximately seven km long and braided lower Tana River (*Finnish:* Teno, papers II and III) was selected due to its almost natural state (only some bank protections exist: Alaraudanjoki, 2001), making it an ideal environment to study the effects of different magnitude discharges on river channel evolution (Fig. 4). The 400–900 m wide braided reach lies upstream of the tidal limit (Collinson, 1970). The Tana River drainage basin measures 16 000 km² of which 11 000 km² lies in Norway. The river itself belongs to the north boreal/subarctic climate with short, cool summers (Kersalo & Pirinen, 2009). The average temperature in July is 12.5 °C (Dankers, 2002; Drebs et al., 2002). By comparison, in January it measures approximately -12.4 °C. In the area, the annual precipitation may be slightly lower (200–700 mm/y) than the average precipitation in Finland. The base level fall is 1–2 mm per year (Sørensen et al., 1987). Most of the annual run-off (65 %) occurs during spring in the May–June period (Dankers 2002). The largest recorded flood has occurred on the 21st of May 1920 at the Polmak observation station with a flow rate of 3844 m^3/s (Alaraudanjoki et al. 2001; Dankers 2002). According to observations by the NVE (Norwegian Water Resources and Energy Directorate), other great flood discharges exceeding 2000 m^3/s have occurred in 1917, 1927, 1932, 1968, 1978, 1984 and 1987. Korhonen & Kuusisto (2010) studied the discharge changes of one tributary of Tana River (Utsjoki) 1963–2004 and found a 0.1 % statistically significant decreasing trend (-9.4 % per ten years) in low discharges. During the study period 2008–2009, the spring snow melt discharges measured approximately the yearly average high discharge (1600 m^3/s) in the Tana River, while the summer and autumn discharges of 2008 were of typical magnitude (paper III; *cf.* Alaraudanjoki et al., 2001).

The Kokemäenjoki River (papers I and IV) represents an outlet river of the lake area and runs through the city of Pori into the Baltic Sea (Fig. 5). It belongs to the South Boreal climate with its favourable summers (Kersalo & Pirinen, 2009). The main stream is 121 km long and its watershed (27 000 km²) is the fourth biggest in Finland (Koskinen 2006). The river is heavily regulated and at present the greatest discharges are experienced during the spring and autumn. The lower Kokemäenjoki River (~40 km), which was selected for the present study, is one of the greatest flood hazard areas in Finland. Prolonged precipitation in the upstream watershed, snow melt, ice dams and frazil ice are the major causes of flooding around the city of Pori (Timonen et al., 2003). The greatest flood, i.e. an approximate 250-year flood, occurred in spring 1899, when over 47 500 hectares of land were inundated (Koskinen, 2006). Major spring floods have also occurred in 1924, 1936, 1951. In addition, the great winter floods, including ice dam and frazil ice floods, have occurred several times on the lower Kokemäenjoki River: in 1920's, 1974–75, 1981–82, 1982–83 and 2004–2005. Korhonen & Kuusisto (2010) found statistically (5 %) significant 5.9 % per ten years increase in mean winter (December–February) discharges in the Kokemäenjoki River during the period 1932–2004. Therefore, this area is important for studying the impacts of possibly changing future discharges on river channel and inundation areas. It is also expected that flood sensitivity will be influenced by the isostatic land uplift (Laine, 1981). Both the changing global mean sea level and isostatic uplift affect sea level on the Finnish coastline (Johansson et al., 2004; Miettinen et al., 2007). The relative land uplift (i.e. the local uplift rate minus the global sea level rise) has been approximated at 6.07 mm/y (±0.44 mm) outside Pori harbour (Johansson et al., 2004). The most recent studies of Baltic Sea water levels have shown that in the future, the eustatic sea level rise could be greater than earlier measured, i.e. 1.2 mm/y (Ekman & Mäkinen, 1996; Johansson & Kahma, 2010).

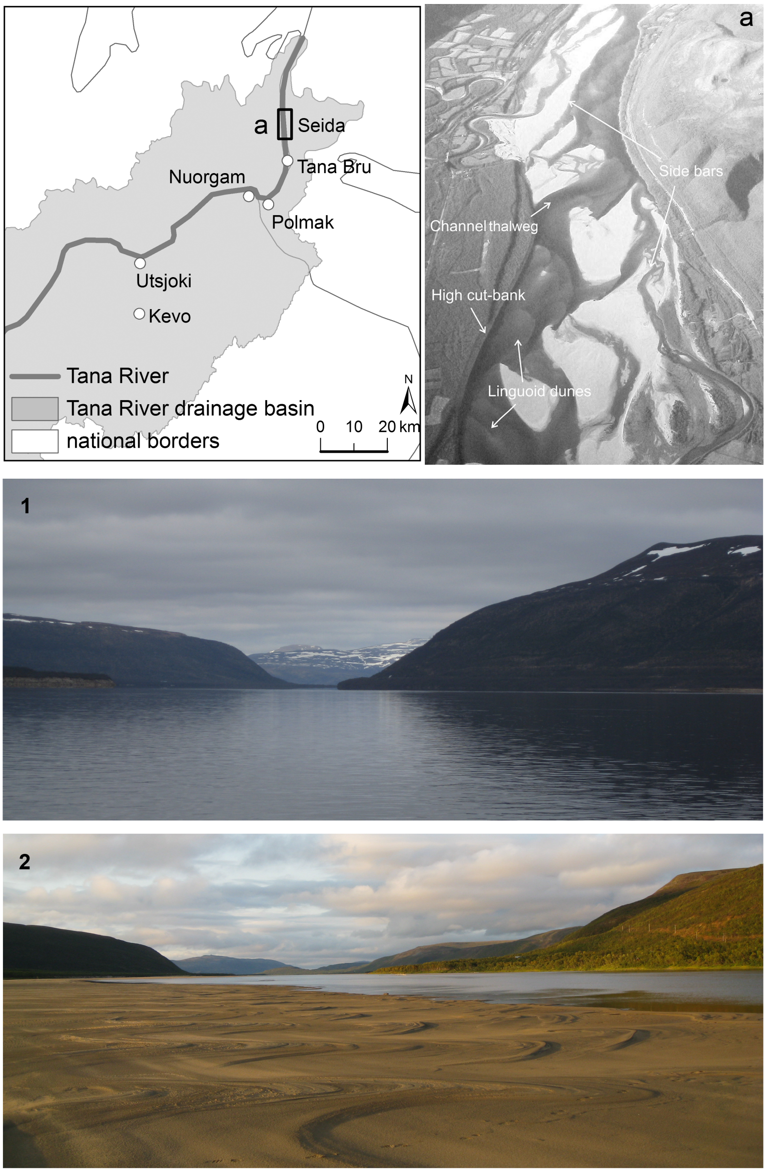

Fig. 4. The Tana River study site (example study site a in Fig. 3) is located in the vicinity of Seida village, northern Norway. The study area is approximately 7 km long. The aerial photograph of the study site was taken on the 8th of August 2004 (photograph by Geoffrey Corner). Sand bars and dunes are widespread, and the differences between seasons are remarkable: 1) Tana River during the spring discharge peak 2010, 2) and in August 2008.

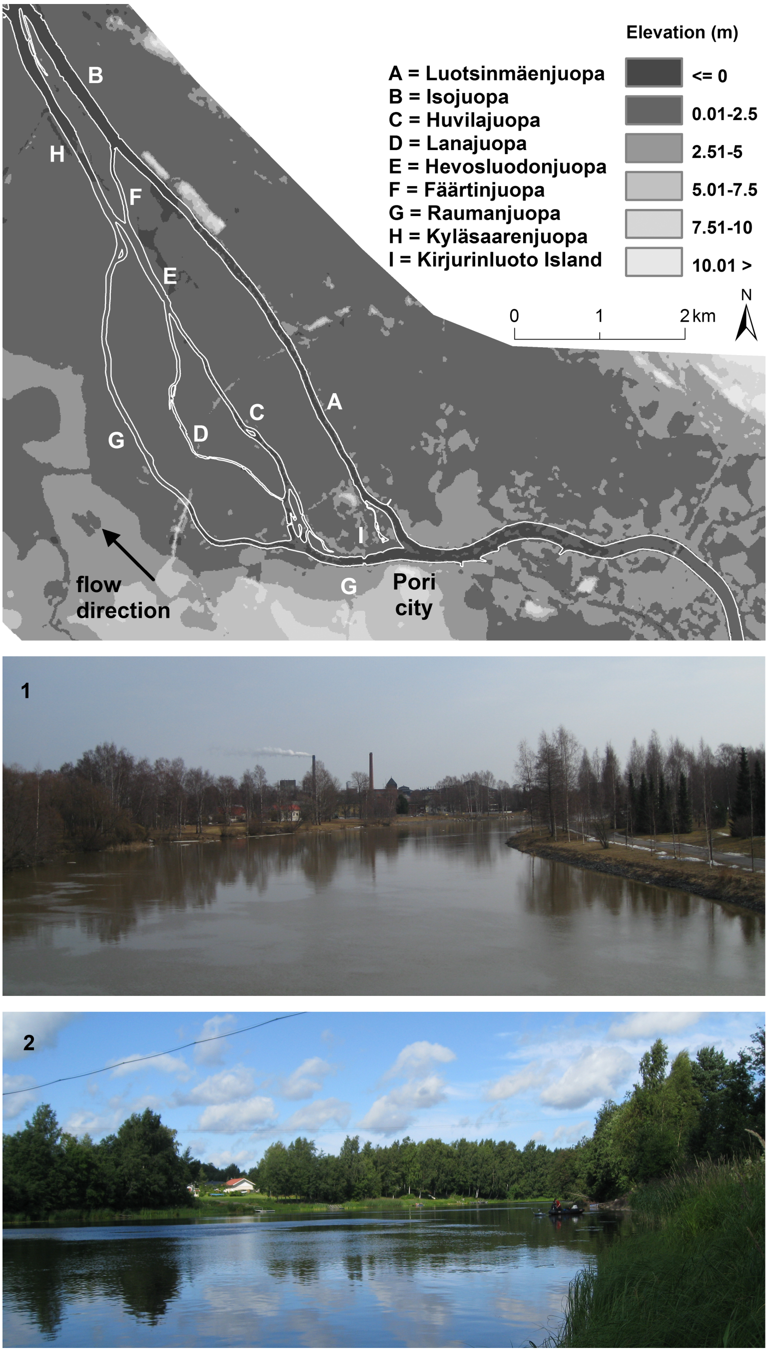

Fig. 5. The lower Kokemäenjoki River study site near the city of Pori (example study site d in Fig. 2). The channel has divided into distributaries (*Fin*. juopa) in the lower reach downstream from the city. During the spring snow melt discharge peak, the water is usually high close to Pori city (1: April 2010). The banks are vegetated during summer (2: July 2009).

3.2 Sedimentary conditions and channel characteristics

Soils and sediment available for fluvial transport also differ in coastal, central and northern Finland. Soils may be broadly classified into bedrock (bare land or deposits less than 1 m), till (and till formations), gravel- and sand deposits (esker, delta, marginal and shore deposits), clay and silt deposits (sea- and lake deposits), fluvial deposits (gravel, sand, silt) and peat deposits (Haavisto-Hyvärinen & Kutvonen, 2007). In Finland, soil deposits formed mainly during and after the last glaciation. Till is the most common deposit type in Finland. Fluvial deposits form only 1 % of the total area of the country, while clay sized sediments cover 8 % and are most abundant in the flat areas of the coast, particularly in south west, where they may be up to 100 m in thickness. In the hilly north, deposits are mainly till, which is also found in the lake-dominated areas of forested central Finland.

The reaches of the Tana and Kokemäenjoki rivers selected for more detailed studies represent the different types of loose boundary rivers (papers II–IV). The lower Tana River represents a braided sand-gravel river, where most of the sediment is transported as bed load. A major proportion of the flood deposits in the lower Tana reach have a median grain size of 0.2 mm and the dominant sediment in the river valley is a well-sorted fine sand of fluvioglacial origin (Mansikkaniemi 1967, 1970, 1972). The lower Tana River runs in a former fjord through late and post-glacial valley fill sediments, which are up to 30 m thick around the study area (Eilertsen & Corner, 2011). Although, the river banks are high and the study area is laterally stable, some erosion is taking place on the left bank. River erosion and evolution reflect the seasons in regions dominated by snow melt discharge (Koutaniemi, 1985). The braid intensity is particularly high during the low water period and the channel has a slight tendency to meander (Collinson, 1970). Linguoid dunes, side bars and ripples are present. Material is mainly transported throughout the year as bed load, with a small proportion of suspended load, while the greatest bed load transport is mainly associated with migrating dunes. The large and rather sudden changes in discharge throughout the summer cause the river bed to be mostly in disequilibrium with the flow, with bed forms controlling the flow patterns during the low water periods (Collinson, 1970).

The lower Kokemäenjoki River (papers I and IV) consists mainly of cohesive clayey sediments and similar sedimentary properties also occur elsewhere in this coastal river region. This river has branched into many small distributaries in its estuary. The concentration of total suspended solids has been classified as high throughout the year (Laine, 1981). The cohesive clays are approximately 10 m thick along the river in the city centre of Pori (Soil map, 1984). Coarser particles, and even boulders, occur at both the Arantila and Ruskila rapids. Sedimentation and erosion areas vary along the reach: erosion dominates the section from the Harjavalta dam to Pori city, although deposition is greater at Pori city centre (Häkkilä, 1979; Kirkkala, 1996; Kiirikki et al., 2004). Moreover, due to the strong flow, silting and sedimentation are insignificant in the two main distributaries (Luotsinmäenjuopa and Raumanjuopa) (National Board of Waters, 1981). Sedimentation occurs mainly in Pihlavanlahti Bay at a rate of 3 cm per year (Laine, 1981). The deltaic estuary also acts as a sedimentation area during periods of high discharge and sediment load events (Kiirikki et al., 2004). The salinity of Pihlavanlahti Bay is below the critical level of 1 ppt for significant flocculation to occur (Krone, 1962; Kiirikki et al., 2004).

4 Methods

4.1 Data collection

4.1.1 Bathymetric data

The bathymetry was measured from a boat in both the Kokemäenjoki (2007, paper IV) and Tana Rivers (2008–2009, papers II-III) using point sonar together with RTK-GPS (see details from papers II–IV) (Fig. 6). The bathymetric data was measured in cross-sectional profiles with an along line point density of 2–5 metres, depending on location. The bathymetric data sets were combined with the bank elevation data derived from the 10 m Finnish national digital elevation model (Kokemäenjoki River) or from 1 m interval contour lines derived from the NVE (Tana River).

The point echo-sounded 2003 bathymetry data from the Kokemäenjoki River was also received from Finnish Environment Institute (papers I and IV). The positioning

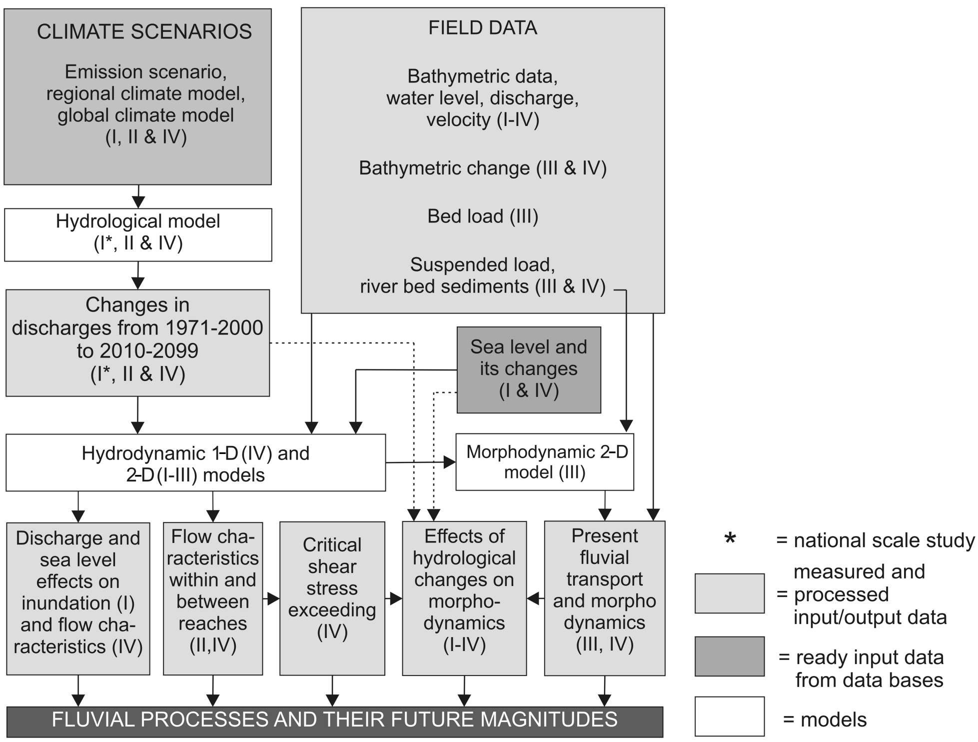

Fig.6. The methodological approaches of this thesis applied for gaining information on fluvial processes and their future magnitudes. The national scale studies are marked with an asterisk. All the other approaches have been performed at the reach scale.

was measured using a real time differentially-corrected GPS. In addition, the 2010 bathymetric data was derived from the Kemijoki Aquatic Technology (KAT) Ltd company, using AquaticSonar swathe sonar (paper IV). This applied interferometric measurement system calculates the depth from the elapsed time and the direction of the sound sent from various transducers. Interferometric systems allow a large number of measurement points and the data from backscattering of the bottom shape may consist of about 40 000 complex samples per one swathe. The channel was scanned by KAT Ltd from the Harjavalta dam to the estuary and the inertia and location (VRS-GPS) were measured simultaneously. Grids measuring 1 x 1 m of regularly-spaced point data was produced from the original data for all water areas accessible by boat (paper IV).

In addition, for the flood inundation analyses undertaken in Salo, Lapua and Kittilä (paper I), the river channel and flood plain geometries were received from the Finnish Environment Institute, Regional Environment Centres and the Department of Geography, University of Turku (Selin, 2006). The final grids were created from input data sets, which consisted of photogrammetric TIN models in Lapua and Kittilä and a 25 m national digital terrain model (DTM) in Salo. Digital contour lines of either 2.5 or 5 m intervals were also used in Salo and Kittilä to interpolate the DTM grid. Further, the DTM in Salo was enhanced with point elevation data measured by the University of Turku, Southwest Finland Regional Environment Centre and the City of Salo (Selin, 2006). The river channel data consisted of sonar point data measured in cross-sections or longitudinal lines. To minimise the interpolation errors, the channel was first interpolated separately before immersing it to the DTM of the surrounding areas.

The compiled geometry data were converted to grid form with a resolution of 10 m for 2-D simulations (papers I–III). The bathymetric data of the Kokemäenjoki River (paper IV) was converted into a Triangulated Irregular Network (TIN) and cross-sections were extracted for 1-D hydrodynamic simulations. These geometries were exported to the various hydrodynamic and morphodynamic models. These applied models, which incorporated finite difference methods, were the 2-D TUFLOW (papers I-III), the 2-D TUFLOW-MORPH (paper III), as well as the 1-D HEC-RAS (paper IV).

4.1.2 Discharge and water level measurements

Discharges were derived from 67 observation stations from the FEI's database to make a national scale assessment (paper I). These sites had at least 29 years of daily discharge observations with over 60 % having observations spanning at least 50 years (mean 67 years). These data were not only applied to calibrate the hydrological model, but also the hydrodynamic model at four flood inundation study sites. For Kokemäenjoki River (paper IV), the discharge data from the FEI were also applied: the upstream discharges from the Harjavalta dam were used to calibrate the hydrodynamic model. Instead for Tana River (papers II and III), the discharge data from the NVE's observation station at Polmak were applied. Discharges were also measured using Acoustic Doppler Current

Profiler (ADCP) during 2007–2010 (papers II–IV). The discharges were measured not only at the up- and downstream boundaries, but also from the middle sections. The tributary and distributary discharges were also measured in the Tana and Kokemäenjoki rivers, respectively.

Water levels were derived from the national database of FEI and from the Regional Environment Centres for the flood inundation study (paper I). Water levels for paper IV were also partly derived from the national database and from the municipality of Nakkila. The data was measured hourly or every 30 minutes, depending on location. In addition, six water depth (i.e. water level) sensors and one barometric sensor were installed along the lower reach of the Kokemäenjoki River. At the Tana River (papers II and III), the depth sensors were installed at the upstream and downstream boundaries of the study site. The depth measurements were performed every 15–30 minutes depending on the location and the measurement year. The water levels were measured with RTK-GPS at the sensor locations, and depths were related to these water levels for gaining water level records. Water levels were also measured in both the Tana and Kokemäenjoki rivers during echo-sounding.

The measured discharge data were applied as the upstream boundary conditions for the calibration of 1- and 2-D hydrodynamic models (papers II–IV), as well as for the morphodynamic simulations in paper III. The water levels were applied as the downstream boundary conditions for the hydro- and morphodynamic models.

4.1.3 River bed sediments and sediment transport

Bed sediments were sampled using a van Veen- type bed sampler in both the Tana and Kokemäenjoki rivers during the low summer/autumn discharge periods of 2008 and 2009 (papers III and IV). Samples were also taken from sub-aerially exposed bars on the Tana River. In total, 36 and 107 bed sediment samples (weighing 0.5–1.0 kg) were recovered from the simulation areas of Tana River (2008 and 2009) and Kokemäenjoki River (2008), respectively. The Tana River samples were taken both years from the same locations. Although all of the Tana River samples were dry-sieved, most of the Kokemäenjoki River samples were constituted mainly of very fine sand, cohesive silt and clay sized particles on the Udden-Wentworth scale. These were analysed using a Coulter counter LS200, i.e. a laser diffraction method (cf. McManus, 1988). The D_{10}, D_{50} and D_{90} grain sizes were determined from the samples (papers III and IV). These grain sizes were applied in both the morphodynamic simulation (paper III) and in the determination of erosion and sedimentation potentials (paper IV).

In Kokemäenjoki River (paper IV), the critical thresholds of bed shear stress were defined for the D_{10}, D_{50} and D_{90} grain sizes of each sample. For coarser D_{50} particle sizes (e.g. 0.25–5 mm), the critical bed shear stress or permissible tractive force (Chow, 1959) was determined based on the graph presented in Chow (*loc. cit.:* Fig. 7-10). For small, i.e. 0.004–0.25 mm, D_{10}, D_{50} and D_{90} particle sizes the critical thresholds were determined from the curve defined by Lick et al. (2004) (van Rijn, 2007: Figure 1). Since Lick et al. (*loc. cit.*) used a somewhat different bed shear stress calculation procedure, these values were scaled to the corresponding levels of

Chow (1959) based on the critical value of the 0.25 mm particle size, which both had defined in their threshold curves.

In the Tana River, the bed load was measured (n=54) using a hand-held Helley-Smith sampler (see details from paper III), which has been widely used in different environments (Ashworth & Ferguson, 1986; Warburton, 1992). Further, the bed load measurements were performed at different discharges, water depths and velocities at the upstream boundary and around one upstream bar. The disturbance of bed material due to the sampler touching the channel bed was excluded from the bed load calculation and the grain size distribution was determined after dry-sieving. While bed load measurements were applied as indicative of the bed load, they were also used as a comparison to the simulated bed load amounts.

The total suspended solids (TSS) were measured from cross-sections along both the Tana (2008, paper III) and Kokemäenjoki (2009–2010, paper IV) rivers using an Eijkelkamp sampler. Suspended sediments were also measured using a depth-integrated water sampler along the bed load measurements of Tana River in 2010. In addition, an automatic water sampler (AWS, model: Liquiport2000) was installed in the Kokemäenjoki River. A total of 360 samples were taken in the period April–October 2009. In spring 2010, AWS was installed for the spring high discharge period. Water samples were pump-filtered using 0.7 µm filters (Whatman GF/F) and TSS (mg/l) was calculated. Measurements for total suspended solids were also available for the Kokemäenjoki River from the database of FEI, whose sample processing procedures for the period 1995-2009 were similar to this study.

4.2 Hydrological simulations

4.2.1 Simulations using Watershed Simulation and Forecasting System

The hydrological scenarios of the daily discharges were simulated using the conceptual hydrological model known as the Watershed Simulation and Forecasting System (WSFS) (papers I, II and IV). Although the basic structure of the WSFS is the same as in the HBV model, differences appear in the river routing, watershed description and some process models, such as the snow model (Vehviläinen et al., 2005). The WSFS has been developed into a semi-distributed model since it is divided into over 6000 small lumped sub-basins (40–500 km^2), each of which have their own water balance simulations, and all 2600 lakes in Finland (> 100 ha) are included in the model (Vehviläinen et al., 2005).

The WSFS was calibrated using observations of the snow water equivalent, as well as water levels and discharges from 1980/1981–2008, depending on the paper. Lake regulation, which particularly influences the discharges in the lake area, was modelled with operating rules, where a certain daily water level corresponds to a certain outflow (papers I and IV). Although this approach reproduces the actual regulation well, on average, since the same rule is applied every year, the regulation is not optimal in all individual years. The operation rules were modified to function properly in climate change simulations with a changing timing for spring floods.

The daily precipitation and temperature of climate scenarios (i.e. the climate model, together with the emission scenario) from different RCMs and GCMs were used. These were derived from the ENSEMBLES data archive (Hewitt, 2005). The IPCC (Intergovernmental Panel on Climate Change) SRES scenarios were employed (IPPC, 2000). The climate change signal was transferred to the hydrological model using two methods: 1) direct bias-corrected daily RCM (regional climate model) data as input for the hydrological model (paper IV), 2) the delta change approach (papers I, II and IV). Due to the systematic biases in the RCM outputs, bias correction was necessary for matching the simulation to the observed hydrology. Although several bias correction methods have been proposed (e.g. Graham et al., 2007; Lenderink et al., 2007; Seguí et al., 2010), the quantile-quantile (q-q) method was selected (Déqué, 2007; Seguí et al., 2010; Wood et al., 2004). The q-q method is a relatively simple but effective method that reproduces the observed discharges in the control period (Veijalainen et al., 2012). The simulated distributions (cumulative density functions, CDFs) of air temperature and precipitation in a 0.25 degree grid were corrected to match their observed distributions. When operating outside the 99 percent range, the value of the closest quantile was used. Observations from the period 1961–2000 were used as the control period values (paper IV).

In the delta change approach (papers I, II and IV), a temperature-dependent temperature (tdt) factor was applied for changing the temperatures (Andréasson et al., 2004; Veijalainen & Vehvilainen 2008). Based on the RCM temperatures, the tdt factor increases the lower temperatures during winter, spring and autumn more than warmer temperatures in the same periods (Andréasson et al., 2004). As a result, the temperatures around 0 °C do not increase as much as those around –20 °C, although the temperature changes were scaled so that the average monthly temperatures (of each month) changed by the amount given by the climate scenario. Precipitation values were altered with the direct delta change method, which assumes that the number of precipitation days does not change and that all precipitation days for the respective month change by the same percentage. The years spanning 1971–2000 were applied as the reference (i.e. control) period.

In total, 20 future climate scenarios were applied in paper I, eight in paper II and four in paper IV to produce future discharge scenarios. These hydrological scenarios were simulated for both the reference periods and for future 2010–2039 (paper I) and 2070–2099 periods (papers I, II and IV).

4.2.2 Frequency analysis

The 100-year discharges were calculated in papers I and IV. In paper II, 2-year discharges were determined for depicting the more frequently occurring discharge events and 250-year discharges for more rare events. The maximum discharges for each year were selected from 30 years of daily data. The hydrological year from September to August was applied, since in the future the long-lasting floods of the large watersheds often occur during winter. An extreme value type I (e.g. the Gumbel) distribution was fitted to the yearly maximum discharges for each period, using the method of moments (based on Kite 1977: see paper II for details). The Gumbel distribution was chosen since it is most commonly used and recommended by the dam safety guidelines in Finland (Ministry of Agriculture and Forestry, 1997). Estimates were calculated based on the fitted distribution, particularly for the reference period and future 2-, 100- and 250-year flood discharges.

4.3 Simulations of hydro- and morphodynamics

4.3.1 Hydrodynamic simulations and flow characteristic calculations

Using 10 x 10 m grid geometries, the hydrodynamic 2-D model (TUFLOW) was applied to calculate the flood inundation area (paper I), to determine the characteristics of future flow (paper II) and to calibrate the hydrodynamics of the morphodynamic model (paper III). The 1-D hydrodynamic model (HEC-RAS) was also applied to calculate future flow characteristics (paper IV). The geometry was imported to the 1-D hydrodynamic model as cross-sections.

The hydraulic models were calibrated by adjusting Manning's n values of bed roughness so that the simulated water levels matched those observed (papers I–IV). After calibration, the flood inundation extents of the 100-year discharges were simulated for selected four study sites representing different watershed categories as steady flow 2-D simulations (paper I). The results of the hydrological modelling, i.e. the minimum and maximum 100-year discharge in both the 2070–2099 and reference periods, were used as the upstream boundary conditions, while the corresponding water levels were used as downstream boundary conditions. Different sea water levels depicting the commonly observed and extreme situations were applied as the downstream boundary conditions in the coastal area reaches (Salo and Pori).

In addition, the channel modification potential of the frequent (2-year) and infrequent (250-year) future floods was examined in Tana River by simulating them as unsteady flow (2-D: paper II). The spatial variation of velocity, bed shear stress and stream power were determined from these simulation results. The channel width was at least 10–20 times greater than depth in most of the cross-sections and therefore, it was possible to use flow depth as a substitute for the hydraulic radius in bed shear stress calculations (Chow, 1959; Magilligan, 1992; Rhoads, 1987). The bed shear stress was calculated as a function of the unit weight of water (9800 N/m^3, Eq. 1) (Aggett & Wilson, 2009; Baker & Costa, 1987; Ferguson, 2005), depth (D, m) and water surface slope (S_w, m/m). The bed shear stress was then calibrated to a water temperature of +5°C using the following formula:

$$\tau 5 = \tau_{bed}(\mu 5/\mu)^{2/3} \qquad (12)$$

where μ5 and μ are the water viscosities at +5°C (1.519×10^{-3} Pa·s) and at +25°C (8.91×10^{-4} Pa·s), respectively. The stream power per unit area was then calculated based on the calibrated bed shear stress and flow velocity (*v*) for each modelled cell (Eq. 4).

Unsteady 1-D simulations of future discharges were also performed in Kokemäenjoki River for future shear stress determinations (paper IV). The simulations of 30 years were performed for both the reference and 2070–2099 periods by applying four selected scenarios. The future unsteady flow simulations were first performed without future sea water level changes. Subsequently, simulations were performed which applied two discharge scenarios with the delta change method and the two latest approximated Baltic Sea water elevation scenarios (Johansson and Kahma, 2010). The boundary shear stress was calculated at each cross-section of the 1-D hydrodynamic simulations using Eq. 2, where the slope of the energy grade line (m/m) was used. The 30-year daily average values of boundary shear stress were calculated for all control period and future unsteady flow simulations. The flow parameter changes in different seasons were detected and compared to the control period. According to Partheniades (1965), the minimum scouring shear stress for dense and deposited cohesive bed material is approximately 0.057 N/m^2. Comparisons were made against this scouring shear stress, since it was similarly defined as boundary shear stress in the HEC-RAS.

In Kokemäenjoki River, in addition to the simulations of the 30 years long periods, the 100-year flood discharges (of paper I) and the average discharges for both the control and future 2070–2099 periods were simulated as steady flow (paper IV). The corresponding average water levels were applied. The bed shear stresses were calculated based on Eq. 1, using the water surface slope, and values were extracted to the bed sediment sampling points. The exceeding of critical bed shear stresses was analysed against the critical thresholds, which were defined based on the bed sediment D_{10}, D_{50} and D_{90} values. In addition, those sediment sample locations were detected where the critical shear stress value of 1.7 N/m^2 (Williams, 1983) was exceeded. According to Williams (*loc. cit.*), this particular bed shear stress value moves particles up to 10 mm in size.

4.3.2 Morphodynamic simulations

The morphodynamic simulations were performed in Tana River using the TUFLOW MORPH 2-D model, which was run in a semi-coupled manner with the TUFLOW hydrodynamic model (paper III). This model has been under development since 2004. Sediment transport and bed update components are included in the model. The hydrodynamic engine passes water surface elevations, depths and the velocity for sediment transport rate calculations.

The sediment transport algorithms of Meyer-Peter & Müller (MPM: 1948, Eq. 5) and van Rijn (VR: 1989, Eq. 6) were selected due to their wide acceptance and suitable grain size ranges. Van Rijn's relationships divide transport into bed and suspended load, which occurs when the shear velocity exceeds the fall velocity. Changes in viscosity caused by temperature were considered by altering the fall velocity in the VR simulations. Fall velocity was set to correspond to the average measured temperature (9 ºC) level.

The effects of longitudinal (attributed to Bagnold, 1966 in Van Rijn, 1993) and transverse bed slope (method of Ikeda, 1982, as advocated in Van Rijn, 1993; Lesser et al, 2004) have been incorporated into the model. An upwinded numerical scheme was adopted, which has been found to be extremely stable (Lesser et al., 2004). In this case it was applied to a rectilinear (as opposed to curvilinear) grid and sediment transport rates were calculated at the centre of each computational cell. Although a four point averaging filter was applied to the initially calculated bed change values, averaging was instead applied by weighting, i.e. in a non-uniform manner, in accordance with the amount of bed change calculated prior to filtering. In this way, the occurrence of unrealistic changes was prevented in areas away from morphological activity. The effects of helical flow, and a bank collapse mechanism were applied during the bed update calculations and the bed update scheme solved the hyperbolic sediment conservation law, i.e. the Exner equation (Bernini et al., 2006; Doeschl et al., 2006; Nicholas et al., 2006). The sediment transport (ST) scaling was tested with different factors ranging from 0.5–2 (van Rijn, 1984a) in both the VR and MPM simulations. In addition, both spatially varying and constant D_{50} (VR and MPM) and D_{90} (VR only) grain sizes were applied in simulation tests. In total, eight final year-long (27.5.2008–24.5.2009) simulations were performed with different algorithm, grain size and ST scaling combinations.

The modelled river bed elevations, bed load, flow velocities and depths during spring floods, secondary discharge peaks and lower discharges were analysed against the bed load, bathymetric and ADCP measurements of 2008–2009. Volumetric changes were compared to the observed one-year changes. The best simulation results were selected for analysis of the effects of different magnitude discharges on river bed evolution. Volumetric changes were calculated for certain time periods representing different discharge conditions. The channel development during one year were also analysed by plotting the cumulative widths against depths and velocities (Nicholas, 2003; Mosley, 1982, 1983) and by calculating averaged braid counts (Mosley, 1982; Egozi & Ashmore, 2009).

5 Results and discussion

The major findings of the present and future fluvial process magnitudes are presented and discussed on both a national (paper I) and local scale between different loose boundary rivers (Knight, 1989) (papers I–IV). The major emphasis is on the detection of spatial, as well as temporal variations in potential future changes between different river reaches (papers I, II and IV). The results are reflected against earlier findings of climate change impacts on discharges, inundation, flow characteristics and fluvial transport potential. Thereby, this work seeks to provide reinforcement for the possible patterns of future fluvial process magnitude changes and channel evolution. In particular, an examination of the relative importance of different magnitude discharge events as present and future channel modifiers is performed (papers I–IV). These findings are applicable for further river environment change detection and flood hazard and risk assessments.

5.1 Present day fluvial transport and morphodynamics

Both low and high discharge conditions greatly influence sediment transport and morphodynamics (papers III and IV). Changes of different magnitude in the river bed, as well as fluvial transport were measured between the Tana and Kokemäenjoki rivers, which represent the two different kinds of loose boundary rivers defined by Knight (1989). One important characteristic of these rivers is their channel material, which causes differences in the modes of transport, as well as morphodynamic changes. The nature of the sediments in the Kokemäenjoki River (paper IV) is mainly cohesive, as 62 % of the samples measured in 2008 had cohesive-type transport properties, with more than 15 % of the mixture of particles less than 0.0625 mm (Mitchener & Torfs, 1996). Similar rivers, where cohesive sediments are present, can be found for example from the coastal hydrological region of Finland. By contrast, the D_{50} and D_{90} grain sizes of the Tana River were measured in 2008 at 0.542 and 0.947 mm, respectively (paper III). These grain sizes are larger than the earlier average estimation of 0.2 mm in the same reach (Mansikkaniemi, 1972). Although the D_{50} grain size in the Tana River is comparable with other braided river reaches, slightly finer material (e.g. 0.22–0.44 mm) has also been observed in braided rivers, such as in the South Saskatchewan River (Sambrook Smith, 2006).

The transport of suspended sediments was measured to be greater in the lower Kokemäenjoki (paper IV) than in the lower Tana River (paper III). This corresponds to the present knowledge of the transport modes in these types of rivers. According to Gao & Josefson (2012), sediment in the Oneida Creek watershed (central New York) is mainly transported by relatively small frequent discharges, with a half-year recurrence interval. This finding that frequently occurring discharges are important transport agents is supported by the measurements from the Kokemäenjoki River. The highest over 100 mg/l suspended sediment concentrations of the Kokemäenjoki River (paper IV) did not always occur during the greatest discharges (~503–628 m^3/s) in the period 1995–2010. These greatest discharges of that period were close to the calculated 2-year discharges (592 m^3/s) based on the Gumbel distribution used for paper I. However, similar high suspended sediment concentrations were also measured with markedly lower discharges, e.g. 360 m^3/s (36 kg/s within cross-section, paper IV). In addition, rather low concentrations (24 mg/l) were measured both during 230 m^3/s and high 610 m^3/s (~15 kg/s within cross-section) discharges. Thus, in some occations the transported amount was even more during rather low discharges (paper IV). While the highest concentrations occurred in 1995–2010 mainly during spring (March–May), they were also high in winter and autumn. Compared to the Kokemäenjoki, the concentrations of suspended sediment are lower based on measurements in the sandy lower Tana River (paper III), although annual variations during 2008–2010 were notable. The suspended sediment concentration immediately prior to the 2008 spring snow melt discharge peak was approximately four times greater than during the lowest autumn water period. However, concentrations similar to the low water period were also measured during the spring of 2010. Bed load domination, which is typical to braided rivers (Carson, 1984; Summerfield, 1991; Murray & Paola, 1994), also occurs in the lower Tana River. In terms of bed load, one interesting finding is that despite a diminishing discharge from 1342 to 550 m^3/s during 2010 spring, the average load decreased only slightly. This phenomenon needs further research with additional measurements.

Field observations show that river bed elevation changes are faster in the lower Tana River within one year than in the lower Kokemäenjoki River over a period of 3–7 years. Although erosion dominates in the main channel and the two largest distributaries of the Kokemäenjoki River, at most, it was only approximately 0.3 m during 2003–2010 (paper IV). During the period 2008–2009, the bed elevation changes of the lower Tana River were mostly less than ±1 m (paper III), which is in accordance with a study by Kiss & Sipos (2007) in the Maros River, Hungary. In addition, the present work gives confirmation for a possible downstream thalweg shifts greater than 100 m in braided rivers within a one year time span. Similar movement has also been observed earlier in the South Saskatchewan River (Sambrook Smith et al., 2006). Thus, it is confirmed that local characteristics affect the geomorphic responses to discharge events.

Based on paper III, 2-D simulations are stated to be applicable in large braided rivers at the macro-form scale. The morphodynamic simulations show that river bed modifications can be fast throughout the channel during the annual flood peaks, while sediment transport and channel modifications are constrained to the braid channels during the lower water periods. Nevertherless, the cumulative river bed changes were greater during low water period of 5.7.–3.9.2008, which were of typical magnitude, than during spring discharge peaks of 2008 and 2009. Therefore, the results support the statement that the cumulative effect of low discharges may cause greater channel modifications than single discharge peaks (Leopold et al., 1964). This emphasises the importance of fluvial transport and channel modifications during the frequent typical low discharges. Despite the different magnitudes of the changes and transportation quantities between the Tana and Kokemäenjoki rivers, the importance of these frequently occurring rather low discharges as transport agents is similarly noticed in Kokemäenjoki River based on 1995–2010 suspended load examinations.

It is noteworthy that even though the 2009 spring discharge of the lower Tana River was 1.5 times (560 m^3/s) higher than in 2008 spring peak, within two days the greater peak discharge only produced of 50–80 % of the changes that occurred during the four days of 2008 spring peak, depending on the simulation used (paper III). The volumetric changes during the spring discharge peaks of 2008 and 2009, which were close to the 1-year and 2-year discharges, respectively (paper II), were broadly similar (paper III). Thus, the results suggest that there is not clear linear relationship between these bed modifications and discharge changes. Since the 1- and 2-year discharges caused almost similar morphodynamic changes, this supports the earlier findings that 1.2–1.45-year discharges may cause the most bed load transport over time (Pickup, 1976). At any rate, further analysis with higher or extreme discharge situations is recommended for lower Tana River and other sandy river reaches.

The study done in the lower Tana River (paper III) confirms that at low water periods, this sandy braided river cannot be considered a very unstable single-thread channel, such as the braided gravel bed river studied by Egozi & Ashmore (2009). During the lowest 2008 summer discharges, the river bed evolved not only in the main braid channel but in other braid channels also. Conversely, it can be stated based on paper III that during discharges greater than 497 m^3/s, the lower Tana River resembles a single-thread channel, i.e. the depth increases and the proportion of shallow areas diminish together with increasing discharges (Mosley, 1982).

It must be remembered that it was impossible to collect winter field measurements in either the Tana or Kokemäenjoki rivers and thus, the winter discharge, sediment transport and morphodynamic change data were based on information from simulations and other databases. Therefore, more detection on fluvial process magnitudes during winter periods is needed.

5.2 Discharge changes

Based on the simulations of this study, there is, depending on the scenario, great variation in the future high discharge change forecasts between rivers and seasons (papers I, II and IV). Although the 100-year discharges are forecast to decrease 8–22 % on average in the period 2070–2099 (20 hydrological scenarios), compared to the reference period 1971–2000, there appears great differences between the applied scenarios (paper I). The possible change in these high discharges at the 67 sites of paper I varies even from -70 to +40 % between the minimum (i.e. the scenario producing the smallest floods in future for each site, different scenario on different sites) and maximum scenarios. This is similar to the work by Arnell & Reynard (1996), who have also found that different catchments respond differently to the same scenario. When compared to the contradictory continental scale studies of Lehner et al. (2006), Graham et al. (2007) and Dankers & Feyen (2008), the results of this work are more similar to the latter study, in which flood decreases have been found in Finland. Thus, if future flood hazard or risk is detected based only on average or region wide high discharge changes without hydrodynamic simulations, which has been common elsewhere (Lehner et al., 2006; Schmidt-Thomé et al., 2006), the potential hazard or risk seems to decline.

The specific regions where an increase or decrease in high discharges (2-, 100-, 250-year) is expected to occur, are definable in Finland (papers I, II and IV). Based on many applied scenarios, increases will be particularly encountered in the large lakes of central Finland, and their outflow rivers (paper I). The Kokemäenjoki River is one of these outflow rivers, where increasing discharges from 1971–2000 to 2070–2099 are forecast (papers I and IV). Simulation results of paper I show that high discharges can also be expected to increase in some small rivers along the southern coast. By contrast, the magnitudes of both the 2- and 100-year discharges are mainly predicted to decrease in the north, as well as most parts of central Finland by 2070–2099, due to increasing temperatures and decreasing snow accumulation. The Tana River is an example of northern rivers (paper II), where these mainly diminishing 2- and 250-year discharges are forecast. One major finding is that the expected change seems to be relatively greater in the more frequent 2-year discharges than in the more extreme 250-year discharges.

The high discharges (100-year) are forecast to diminish on average 15–40% during spring by 2070–2099, whereas increases are expected in other seasons 12–40% (paper I). Even though some scenarios still predict relatively high spring discharges in the north for the period 2070–2099 (paper I), the timing of these spring peaks show a possible change from the 1971–2000 period for example in the Tana River (paper II), occurring two to three weeks earlier. It is also interesting that although spring peak discharge magnitudes are forecast to decrease (paper II), these decreases have not yet been observed in the 20th century (Korhonen & Kuusisto, 2010). These findings give confirmation to the earlier studies of northern Europe, such as Norway and Sweden, where also future spring discharges have been forecast to become earlier and snowmelt floods have been predicted to become more seldom (Beldring et al., 2006; Andéasson et al., 2004; Graham et al., 2007; Beldring et al., 2008).

A common feature for most of the forecasts performed for different areas in Finland is an expected increase in the autumn and winter discharges (papers I, II and IV). Similarly, earlier studies performed in Finland, Sweden and Norway have forecast increasing discharges in autumn and winter (Vehviläinen & Huttunen, 1997; Bergström et al., 2001; Andréasson et al., 2004; Beldring et al., 2006; Beldring et al., 2008). Such different seasonal discharge changes, such as the increasing mean winter (December–February) discharges in the Kokemäenjoki River have already been found in the 20th century (Korhonen & Kuusisto, 2010). These observed trends support the future 2070–2099 forecasts for the Kokemäenjoki River, since due to the increasing precipitation, as well as wetter and milder autumns and winters, autumn and winter discharges are predicted to increase (papers I and IV). In addition, simulations using direct RCM data show that decreasing summer (June-August) discharges can be expected (paper IV). The winter discharges may even become the largest discharge events in some of the outflow and coastal area rivers (paper I). The earlier findings of Vehviläinen & Huttunen (1997) support these forecasts, who also have predicted an increase in winter discharges, particularly in the central lake area. These and earlier findings from other Nordic countries highlight the significance of taking the discharge changes of certain recurrence interval and different seasons into account, when considering riverine planning and flood risk assessments.

Based on simulations, those scenarios with small temperature increases produce the largest future floods in northern Finland, but in the coastal rivers and the lake area largest floods are produced by a scenario with large precipitation increases (paper I). Temperature change is stated to be the most important factor determining the discharge magnitudes of the 2070–2099 period in the northern region, which is dominated by the spring discharge peak at present. Nevertheless, scenarios with large precipitation increases may also cause high discharges in this location (paper II).

Based on paper I, global climate models may cause greater uncertainty than the choice of either the emission scenario or regional climate model. Differences between simulations using GCMs and RCMs are notable for example in Tana River (paper II), while quite small differences can be between different emission scenarios with the same GCM (paper I). Similarly, it has also been noticed by Prudhomme et al. (2003) and Verhaar et al. (2010) that the choice of emission scenario makes much less difference to the results than the choice of global climate model. By constrast, Beldring et al. (2008) have not found clear difference between the results based on different emission scenarios or general circulation models. Conversely, Andréasson et al. (2004) have found that discharge and flow patterns are sensitive to the choice of both the GCM and emission scenario. Thus, based on the results of papers I, II and IV, the application of scenarios from several climate models, especially GCMs, is important for impact studies, a finding supported by Prudhomme et al. (2003).

5.3 Changes in flow characteristics within and between different rivers

The flow characteristics, such as velocity, bed and boundary shear stress and stream power can be expected to follow the future discharge changes (papers II and IV). Earlier, flow characteristics and thus, erosion and sedimentation potential, have also been reported to change similarly along with discharges and water surface slope (Arnauld-Fassetta, 2003; Nicholas, 2003; Verhaar et al., 2011). Simulations show that if the the maximum future 250-year discharge scenario realizes in the lower Tana River, velocity, bed shear stress and stream power increase from the 1971–2000 to 2070–2099. Instead, all other 250-year discharge predictions show reductions in these flow characteristics (paper II). During all of the 2-year discharge events of 2070–2099, velocities, bed shear stresses and stream powers are predicted to be smaller than in 1971–2000. In the case of average stream power, the range of change is almost exactly the same as the range of relative changes in both the 2-year and 250-year discharges. Therefore, it can also be assumed that the average stream power values in other areas may show similar changes to the discharges. Based on the analysis, it can be stated that the reduction of the actual and relative average values of velocity, bed shear stress and stream power and thus, erosion potential, will likely be larger in the more frequent 2-year discharges than in the more infrequent 250-year discharges.

The variation in the flow characteristic magnitude changes are evident within the river reaches studied. The analysis performed in the Tana River show that when compared to the 1971–2000 period, the area percentages for lower velocity (<0.75 m/s), bed shear stress (<5 N/m^2) and stream power (<5 W/m^2) increase with all future 2-year discharge scenarios, including a minimum future 250-year discharge scenario (paper II). Based on the simulations, the spatial variability and reduction in the stream power between the control period and the maximum scenario can be expected to be greater in the frequent 2-year discharges than in the 250-year discharges. In spatial terms, the greatest stream power reduction is forecast for the future 2-year discharge on sand bars and in confluence areas, while the least reduction is predicted to take place near the thalweg areas. Within the lower Kokemäejoki River (paper IV), spatial differences in boundary shear stress and bed shear stress exist. The greatest boundary shear stresses of for example the 100-year discharges locate in the upstream part of the study reach, where the variability in river bed elevation is greatest. In addition based on future simulations, it can be stated that the increasing future discharges increase the bed shear stresses particularly where the variation in both depth and water surface slope are high. However, the analyses show that there is no clear linear downstream lowering trend in the main channel of the Kokemäenjoki River.

This finding is similar to the studies where no clear trend in stream power has been found along the longitudinal profile (Barker et al., 2009; Phillips & Slattery, 2007). It must be remembered that only the effects of future discharge changes on flow characteristics were analysed in papers II and IV, since it was not possible in these simulations to take the actual bed form modifications into account.

While bed shear stress magnitudes and their predicted changes differ between the Tana and Kokemäenjoki rivers, it must be borne in mind that the bed shear stresses in the former reach were calculated by calibrating them to 5°C. Therefore, based on dynamic viscosities of +25 and +5 °C, these calibrated values are 1.4 times greater than without calibration, and the values presented in papers II and IV are not readily comparable between these two areas. In the Tana River, the maximum bed shear stress of the maximum future 250-year discharge is forecast to be approximately 25 N/m^2. Accordingly, higher future bed shear stresses are expected in the Kokemäenjoki River than in the Tana River, since the highest calculated bed shear stresses without calibration range likely from 60–100 N/m^2 with 100-year discharges, depending on the discharge and water level scenario. Further, it was not possible to take the yearly variation of temperature into account in the Kokemäenjoki River (paper IV). Therefore, in reality, even higher future magnitudes of shear stress may be expected during the coldest months than presently shown. Thus, although one of the characteristics of braided rivers has normally been the greater shear stresses than in straight channels (Parker, 1976; Schumm, 1985; Ferguson, 1987; Kleinhans & van den Berg, 2011), it seems that the present and future shear stresses are greater in the straighter river reach of this study. Overall, the calculated shear stresses of the Kokemäenjoki and Tana Rivers are rather low even with infrequent discharges, when compared to the shear stresses of bank-full discharges in Belgium (Petit et al., 2005) or from large flash-floods in the USA (Baker & Costa, 1987).

5.4 The combined effects of discharge and sea level changes on inundation and flow characteristics

The results of papers I and IV show the future projections for flood inundation and flow characteristics in rivers representing the different watershed types in Finland, due partly to changes in discharge and sea level but also to a combination of these factors. Such a simulation result comparison between different river reaches has not been performed previously. Based on the analysis, there is no linear relationship between the changes in discharges and inundation extent (paper I). Similarly, sea water level and the inundation extent may change non-linearly to each other. Nevertheless, increasing future discharges are expected to produce increasing inundation, while decreasing flows may cause a reduction in flooding. This reflects the work of Thonon & Klok (2007), who also have found that the extent of inundation shares the same trend with flooding frequency.

The hydrological simulations (paper I) indicate that the hydrological scenario, which produces the maximum and minimum inundation extents, varies between regions, while the magnitude of change varies between the reaches also. Even rather small discharge (11–32 % reductions) and water level changes may cause changes in the extent of inundation (12–90 % reductions) in a topographically flat study site (e.g. Lapua settlement), and the rate of the simulated inundation change will probably be greater than the discharge change. In one of the snow melt dominated regions of northernmost Finland (Kittilä settlement), the flood inundated area is limited by the surrounding fjells and the future changes in inundation area are all forecast to be rather close to those of the reference period, i.e. between +7 and -18 %. The northern lower Tana River flows in an even more confined channel (paper II), where steep sandy terrace slopes rise 10–20 m above the river (Eilertsen & Corner, 2011). Thus, due to the local characteristics, differences even exist between the forecasts for future inundation extent changes within the same hydrological regions (papers I and II). Instead of performing a flood risk analysis based on only the projected discharge changes, it is therefore recommended whenever possible to employ a numerical simulation, which takes the local topographical characteristics into account.

In low lying coastal area river reaches where the sea acts as the downstream boundary (Uskelanjoki and Kokemäenjoki rivers), the inundation area is greatly affected by sea level (paper I). While the increasing 100-year flood discharges of Kokemäenjoki are projected to cause changes in the inundation extents, the effect of sea level on the inundation is evident in the lowest delta area based on the 2-D hydrodynamic simulations of paper I. The sea level also affects the water surface slope and thus, the water depths, based on the 1-D hydrodynamic simulations in lowest 30 km of the Kokemäenjoki River (paper IV). Similarly, when 1-, 10- or 100-year sea water level scenarios have been analysed on the northern Adriatic coast, the inundated flood risk areas have been also forecast to increase until the year 2100 (Gambolati et al., 2002). Thus, as observed in the cases of both the Uskelanjoki and Kokemäenjoki rivers, the possible future rising sea level may impact strongly on the inundation extents in coastal Finland, while on some occasions sea level impact may even be greater than the effects of the discharge changes.

Increasing future discharges are forecast to increase the erosion potential, while high future sea levels may cause decreases due to the backwater effect of the sea along the lowest 30 km of the Kokemäenjoki River (paper IV). Based on simulations, the greatest fluid forces, and thus erosion potential, occur at high discharges with low water levels. Similarly, Lamb et al. (2012) showed that erosion increases when the flow depth reduces in the river mouth under high discharge conditions. High sea levels will likely result in decreased future daily average boundary shear stresses, whereas the lower sea level scenarios may cause higher shear stresses when compared to “no water level change” situation of a certain discharge scenario (paper IV). In addition, bed shear stresses are forecast to be greatest during the 100-year discharges with the lowest applied sea water level (paper IV). Thus, the sea level may greatly affect the potential for both erosion and deposition also in other coastal area river reaches and estuaries in which future changes in sea level have been forecast (Purvis et al., 2008; Rennie & Hansom, 2011). Therefore, when detecting the effects of future discharge and river channel change on future flood hazard and risk in coastal river reaches, the impacts of sea level should not be ignored.

5.5 The effects of hydrological changes on morphodynamics

The effects of different magnitude present and future discharges on morphodynamics are highly stream-specific, a finding supported by Fuller (2008). Although discharges are forecast to mostly decrease (2070–2099) in the braided lower Tana River (paper II), the velocity, bed shear stress and stream power thresholds defined by Williams (1983) for 10 mm particle movement are predicted to mainly exceed with future frequent 2- and infrequent 250-year discharges. Exceeding of this critical value is also confirmed by the simulations of paper III, since the 2008 and 2009 spring discharge peaks, which were rather similar to the 1- and 2-year discharges, caused rapid and channel-wide volumetric changes. Thus, future 2-year discharges may still cause notable channel modifications in a sandy braided river. Based on paper II, with both GCM and RCM scenarios, the daily average discharges of 30 years are projected to increase from 1971–2000 to 2070–2099 in autumn, winter and earlier in the spring, while the snow melt discharge peak is expected to diminish (paper II, Fig. 7). This indicates that although spring discharges will probably still induce great widespread changes to the river bed, the proportion of the spring flood as the greatest sole bed modifier may decrease and more rapid bed changes likely occur particularly in autumn. With increasing future autumn discharges and a concurrent rise in stream powers, the sediment transport is also expected to increase. Similarly in Denmark, Thodsen et al. (2008) have forecast increasing winter transport. In addition, high magnitude winter discharge events have also been predicted to increase in the Saint Lawrence River, Canada (Verhaar et al., 2011). The importance of the cumulative effects of discharges on the river bed other than during the spring flood season may increase in the Tana River (papers II and III). Since the mean summer discharges appear to remain similar to the present situation (paper II), there will still be a period in the future when modifications are confined to multiple threads, such as observed in 2008–2009 (paper III). Based on papers II and III, it can be assumed that the braided lower Tana River will also greatly evolve in the future, even if the high spring discharges decrease.

Based on this study, overall, the proportion of locations where the critical bed shear stress threshold is exceeded during both the average and 100-year discharges will likely increase along future discharge increases in the clayey Kokemäenjoki River (paper IV). Even though, the future critical bed shear stress threshold exceeding is expected to be rather similar to the average discharge conditions of control period if the 30 year (2070–2099) average discharge of minimum future scenario comes true. Instead, the exceeding of the threshold will approximately double if the 30 year average discharge of the maximum scenario realizes. Simulations show that every future 100-year discharge scenario with a sea level height of 0.4 m likely causes the critical threshold to be exceeded in more than 90 % of the analysed locations. Based on these particle movement forecasts, the net erosion is expected to continue to dominate in most of the study area during future high discharge conditions. Similarly, Verhaar et al. (2011) have forecast larger future flood events, accompanied with an increasing bed material transport and number of transport events for the tributaries of the Saint Lawrence River, Canada.

Despite the increasing winter and autumn boundary shear stresses in the lower

Kokemäenjoki River (paper IV), the proportion of days without movement, i.e. when the minimum scour shear stress (boundary shear stress) of Partheniades (1965: 0.057 N/m^2) is not exceeded, may also increase with the future scenarios. Therefore, in the future the deposition may be greater during other than high discharge seasons. This is comparable to the findings of Thodsen et al. (2008) in Denmark, who have predicted decreasing transport during the summer months. Since it was not possible to include annual variations in water temperature in the simulations, the contrast between the future summer and winter season transport conditions may actually be even greater than shown here.

According to paper IV, the percentage of days when deposition exceeds erosion (i.e. initiation of particle movement) increases in locations closest to the sea, when the predictions of higher future sea level are compared to the ones of "no sea level change". Although more thorough detection of such change also needs further studies in other regions similar to the lower Kokemäenjoki River, the study by Blott et al. (2006) supports this finding of the relatively high importance of sea level on erosion and deposition. Blott et al. (*loc. cit.*) have found the changes in sea level to have a relatively greater effect on estuary morphology than they have had in the past. A fall in the base level, defined as downstream water level, has been found to lead to degradation in the rivers currently experiencing aggradation or in equilibrium, as well as an amplification of the effects of climate change on sediment delivery (Verhaar et al., 2010). This is similar to the results of paper IV, according to which, the decreasing future 2070–2099 sea water level may increase erosion potential, when compared to the possible higher sea water level situations. Therefore, in addition to the changing discharge magnitudes, sea level change is stated to be an important factor for future morphodynamics.

6 Conclusions and recommendations for future work

The future expected changes in discharges, inundation, flow characteristics and morphodynamics were demonstrated with combined field observation and simulation approaches in different rivers. The detailed conclusions are presented in five major themes.

1) The frequent low discharges of typical magnitude may be even more important for fluvial transport and channel modifications than annual discharge peaks particularly in the sandy braided reach, but great transportation can also occur during these low discharges in the river consisting of cohesive material:
    - Despite the rapid and channel-wide changes caused by 1- and 2-year discharges, the results in the braided sandy Tana River support the earlier findings that the cumulative effect of low discharges can be even a greater driver of river bed changes than a single discharge peak. The morphodynamic models are found suitable for this detection.
    - In addition, the suspended sediment transport in the Kokemäenjoki River suggests that frequently occurring rather low discharges are important

transport agents, since even higher suspended sediment transport amounts were measured during some lower discharge occations than during ~2-year discharge events.

2) The future discharge changes vary greatly both temporally and between different river reaches in Finland:
    - High discharge changes: On average, decreasing discharges are expected in Finland by 2070–2099. In northern areas, mainly decreasing frequent (2-year) and infrequent (100- and 250-year) discharges are forecast. By contrast, increasing high discharge magnitudes are predicted for the large lake areas of central Finland and their outflow rivers, such as the Kokemäenjoki River.
    - Temporal changes: Noteworthy is that both autumn and winter discharges are expected to increase in northern Finland, coastal areas, as well as in the lake area and their outflow rivers. Moreover, in northern areas there may be a shift in the timing of the snow melt high discharge event, which may occur a few weeks earlier in the spring. It is important to take such temporal changes in high discharges into account when making flood risk assessments.
    - GCMs are stated to be a greater source of uncertainty than the choice of either emission scenario or RCM.

3) Although the flow characteristics (velocity, bed and boundary shear stress and stream power) and future changes in the erosion-sedimentation potential follow the discharge changes, the changes may be relatively greatest in the case of frequent discharges and non-linear changes within reaches are possible.
    - In Tana River, where the discharges are forecast to decrease by 2070–2099, the actual and relative average values of flow characteristics of frequent discharges are reduced more than of infrequent discharges. Local channel characteristics are important factors affecting the future erosion and sedimentation potential.
    - Spatial variations in the flow characteristics and their future changes occur along both the longitudinal and cross-sectional profiles in Kokemäenjoki and Tana rivers. Although, the increasing future discharges increase the erosion potential throughout the longitudinal profiles, non-linear trends are also possible. The relative within reach change in the erosional forces between the different future discharge scenarios is greatest in the present shallow areas, such as sand bars.

4) The inundation extents change non-linearly with discharge and sea level changes, while the backwater effect of high sea level conditions reduce the magnitudes of the erosional forces:
    - Although the extent of inundation changes along with discharges and sea water levels, these are not linear due to local, and in particular, topographical differences.

- In the river reaches closest to the sea, inundation extents may be even more influenced by sea level changes than changes in discharges.
- The importance of the combined effects of discharge and sea water level on the magnitudes of flow characteristics is emphasised. At high water level conditions, the backwater effect decreases the forces exerted on the river bed. For example, bed shear stresses are forecast to be greatest in lower Kokemäenjoki River during high 100-year discharges with the lowest applied sea water level.

5) Great river channel modifications are still expected in the future both in sandy (Tana) and clayey (Kokemäenjoki) rivers. In addition, the importance of the discharge events as channel modifiers during periods other than the present spring snow melt season is anticipated to increase on a century time-scale, particularly due to the likely increasing autumn discharges:
   - Tana River: Both the future frequent (2-year) and infrequent (250-year) discharge peaks are forecast to move the particles of this braided sandy river, despite a possible reduction in discharges. Although the future (2070–2099) spring discharges are expected to cause marked widespread changes to the river bed, the proportion of the spring flood as the greatest single bed modifier may decrease. More rapid sediment transport and bed changes are expected, particularly in autumn. The importance of the cumulative effects of discharges on the river bed other than during the spring flood season is anticipated to increase.
   - Kokemäenjoki River: The exceeding of particle movement thresholds and thus erosion is expected to increase, under both infrequent (100-year) and average discharge conditions. Despite the increasing winter and autumn boundary shear stresses in this southern outflow river of the central lake area, the proportion of days without movement, particularly if high sea level scenarios realize, may increase with future scenarios. Thus, it is likely that the importance of the autumn and winter discharges and infrequent flood events as sediment transport agents will increase in the Kokemäenjoki River and in other similar reaches.

The 1- and 2-D hydro- and morphodynamic approaches, together with empirical observations, enable the estimations of future potential changes in fluvial process magnitudes. Since both discharge changes and different sea level conditions affect the changes in erosional forces and thus erosion-sedimentation potential, these changes and their hydrodynamic simulations can be considered important to include in future riverine environment planning and flood risk assessments. Similarly, Lane et al. (2007) have shown the inclusion of geomorphological processes to be important especially in areas where aggradation occurs. Further studies are still needed in other rivers of similar hydrological and sediment characteristics to verify these future changes and variations in both the erosion and sedimentation potentials. It must be remembered that it was not possible to incorporate land-use changes in the present study. Different human impacts, together with the local geological and geomorphological conditions in different river reaches have earlier been found to cause variation in the timing and course of channel change between regions (Zawiejska & Wyzga, 2010). According to Pattison & Lane (2011), problems arise in establishing a link between land management and flooding, not only due to scale effects (i.e. local *versus* catchment scale) and unique catchment characteristics but also since land-use effects are not mutually exclusive from climate change impacts. Since it has been considered likely that land-use changes amplify the climate change effects, further research on this combined issue is needed.

In summary, it is advisable to include more uncertainty assessments in future hydro- and morphodynamic studies. Verhaar et al. (2011) suggested the application of multiple climate models in forecasting bed load transport. Based on the simulations presented here, a similar suggestion can be made. Also, it would be essential to perform a 3-D simulation approach for a more detailed analysis of the possible river bed changes. The field methods available at present, such as laser scanning, swathe sonar and ADCP, enable denser observations for the analysis of river bed change, sediment transport and flow characteristic. Additional bed and suspended load measurements are also required for further verification of the findings that in the long run, frequent typical rather low discharges may cause even greater transport and river bed modifications than single higher discharge peaks. For this purpose, the morphodynamic simulations of the infrequent high discharge events in different river reaches would also be of great importance. Although these multidimensional hydro- and morphodynamic simulations are data intensive, even longer continuous observations of fluvial processes and their changes are still needed to perform more detailed analyses of present and future channel modifications.

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