Variability, causes and implications of soil moisture for land degradation and vegetation regeneration in Sudan

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In memory of Suad Khalifa Khogali

Thank you

يَا أُيُّنَّهَا النَّفَسُ الْمُطَمَّتَةُ ارْجِعِي إِلَى رَبِّكَ رَاضِيَةً مُرْضِيَةً فَاذْخْلِي فِي عِبَادِي وَاذْخُلِي جَنِّي.

صدق الله العظيم.
Abstract

Sudan suffers from years of vegetation degradation and is also hit by climate change which has fatal consequences on its fragile economy and the lives of its 41 million people. The vegetation-soil moisture relationships describe the various vegetation patterns which occur in the country. Natural regeneration of Sudan’s vegetation remains the only possible solution for combating this degradation and ultimately contributing to the country’s economic and social stability. In light of these facts the current thesis tries to understand the physical circumstances that impact vegetation regeneration via studying the connections between soil water, erosion and some of the main elements of the hydrological cycle, namely evapotranspiration, temperature and rainfall. The studies of soil moisture and its climatic associations in Sudan are rare especially in the thorough way that is presented here. These results have important food security implications, informing agricultural development, environmental conservation, and water resource planning.

The first stage of this research sought to evaluate the spatial distribution of soil erosion as one of the implications of soil degradation which poses a serious environmental and socioeconomic threat to the environment and to mankind. The developed erosion model used multispectral Advanced Spaceborne Thermal Emission and Reflection Radiometer (ASTER) and Moderate Resolution Imaging Spectroradiometer (MODIS) products from March and December 2006 plus a Shuttle Radar Topography Mission (SRTM) digital elevation model. The results allowed the identification of erosion gullies and subsequent estimation of eroded area. River flow network and slope are identified as more important natural factors, in causing erosion in the Blue Nile region than aspect and elevation.

The second stage concentrated on evaluating evapotranspiration as a direct reflection of the dynamics of soil moisture and hence vegetation regeneration. Both potential and actual evapotranspiration vary from day to day and have a seasonal cycle that determines the rates of vegetative growth and water stress. This thesis, among other, compares three methods for the estimation of daily actual evapotranspiration in Blue Nile. The methods are the remote sensing using the satellite based Surface Energy Balance Algorithm for Land (SEBAL) model with Moderate Resolution Imaging Spectroradiometer (MODIS) satellite data, the modified Thornthwaite water balance method, and the complementary relationship method. A sequence of spatial distribution maps of seasonal actual evapotranspiration are produced. From the maps it was concluded that in the dry season, the spatial distribution pattern is determined by the location, aspect, land use and irrigation activities. In the wet season, the spatial distribution pattern followed that of the rainfall distribution. The seasonal patterns of actual evapotranspiration are a result of the combined effect of rainfall and soil moisture. The seasonal patterns of monthly soil moisture followed those of the monthly rainfall, but with about one month delay in the phase.

The third stage aimed at better understanding the soil moisture variability as a fundamental factor for vegetation regeneration during the period 1965-2005. NCEP/NCAR and PERC reanalysis data are used to study the general trend and to understand the soil moisture, temperature and rainfall relations using Mann Kendall test and geographically weighted regression. To further understand dry and wet variations in terms of regeneration demand, the aridity index is used. The results showed that there are decreasing trends of soil moisture on an annual and seasonal level and that the trend is less dramatic or weaker in the dry season (November-April) than the wet season (May-October). The long-term average of aridity index is affected by the reported decline in rainfall during 1965-1985.
الخلاصة

تمثل العلاقات بين النباتات ومياه التربة أنماط الحياة النباتية المختلفة التي تحدث في البلاد، ولكن السودان يعاني من عدة سنوات من تدهور العطاء النباتي والتغير المناخي الذي له عواقب وخيمة على اقتصاده النهوض و布莱ة من فتح سكانه البالغ عدهم 41 مليون نسمة. وبناءً على التحديات البيئية للتغير المناخي في السودان لا يزال ذلك الزيادة المحتمل لمكافحة هذا التدهور والمساعدة في الاستقرار الاقتصادي والاجتماعي في البلاد.

وعلاء ضوء هذه الحقائق تحاول الأطراف فيما في الطواف المادي التي تثر إلى تحدد العطاء النباتي عن طريق دراسة العلاقات بين مياه التربة وتحريها من ناحية، والعناصر الرئيسية للدورة الهيدرولوجية مثل التربة ودرجات الحرارة وطول الأمطار من ناحية أخرى.

وهكذا تظهر في الدراسات المتعلقة بدراسات مياه التربة واتباعاتها المناخية في السودان، نادرًا، خاصة بالطريقة الشاملة التي أوردها هذا، وذا إن نتائج هذا الأطراف لها انعكاسات هامة فيما يتعلق بالأمر الغذائي والنسيم الزراعي والمحافظة على البيئة وتخطيط الموارد المائية.

تسعى المرحلة الأولى من هذا البحث لتقديم التوزيع المكاني لجذور التربة بإضافة واحد من الآثار المرتبطة بتدوير التربة والتي يشكلون فيهما اجتماعيات واقتصاديات خطيرة على البيئة والبشرية. وتم تطوير نموذج لدراسة تجربة التربة باستخدام مستمر الأطراف المحيطة والمكتنات الناتجة من تحداته. وتقويم تكلفة الأرضية ومساحة الناتجة من استعداد الأرضية، كما توصلت الدراسة إلى أن شبكة النظام البري ونسبة الإحراج من الأطراف وجانب الأرضية في النسب في إلغاء التربة في منطقة النيل الأزرق بشرق السودان.

بينما ركزت المرحلة الثانية على تقييم التربة كانعكس مباشر لديناكيميات مياه التربة وبالتالي تجربة العطاء النباتي. إن كلا من عوامل عرفة التربة والفعالية مختلفة عن اليوم الآخر، إذا أن دورهمهما المستدام في حياة معتادات النمو الخضري وضغط المياه وانعكاس هذه الأطراف - ضمن أمور أخرى - بين ثلاث طرق لتحديد التربة العلوي في النيل الأزرق.

الطريقة الأولى

طريقة الاستشعار عن بعد والأقمار الصناعية باستخدام نموذج يعتمد على معايرة توزيع الطاقة.

الطريقة الثانية

الطريقة المعدلة لمعايرة المياه.

الطريقة الثالثة

الطريقة العلاجية الكاملة بعد في مادة الصحافة بين الاختيار الثلاث، ثم استخدام الطريقة الأولى لإنتاج سلسلة من خرائط التوزيع المكاني للتراب الفعلي، ومن خلال هذه الخرائط تم الوصول إلى أن هناك نسبيًا إعداد جزئي نتائج الأرضية وانشئيات ما. ورأى في موسى الماء الماء ففم التوزيع المكاني تتب توزيع سوق الأرسط وان الأنماط السويدي للتربة الفعلي هي نتيجة للتذبذب المشترك لهطول الأمطار ومواد التربة وتنبؤ الأنماط المحسوبة لفصول الأمطار والمياه التربة لتبث الأنماط المسبقة لفصول الأمطار. ولكن هذا تأخير شهر واحد.

أما المرحلة الثالثة والأخيرية فتهدف إلى فهم أفضل لتوزيع مياه التربة باعتبارها عاملًا أساسيًا لتجديد العطاء النباتي خلال الفترة 1965–2005، وتشمل البيانات المتاحة من إعادة التحليل لدراسات الآجل العام وفيها شروط وعلامات مياه التربة ودرجة الحرارة والأمطار. كما تم استخراج مساحة الجفاف لمجرة الإعالة بين مオスى الماء والأمطار من حيث الحسابات الجغرافية للعازل، وظهرت النتائج أن هناك اتجاهات تشير إلى أن مياه التربة السنوية والمائية أخذت في التدفق خاصا في موسم الأمطار (مايو - أكتوبر)، كما أن متوسط المدة الطويل لمجرة الجفاف يتأثر بتناقص هطول الأمطار.
Sammendrag

Sudan har vært utsatt for flere år med degredasjon av vegetasjonen og er rammet av klimaendringer som har ført til fatale konsekvenser for Sudans skjøre økonomi og livene til landets 41 millioner mennesker. Vegetasjons- og markvannsrelasjoner beskriver de ulike vegetasjonstypemønstre som forekommer i landet. Naturlig regenerering av Sudans vegetasjon er den eneste mulige løsningen for å bekjempe degredasjonen, noe som vil bidra til landets økonomiske og sosiale stabilitet. I lys av disse fakta søker denne avhandlingen å forstå de fysiske forholdene som påvirker vegetasjonsregenerering ved å studere sammenhengene mellom markvann, erosjon og noen av de viktigste elementene i den hydrologiske syklusen, nemlig fordampning, temperatur og nedbør. Undersøkelser av markvann og dets klimatiske avhengigheter er sjeldent i Sudan, spesielt på den grundige måten som er presentert i dette studiet. Disse resultatene har viktige matsikkerhetsimplikasjoner, samt informasjon om relevans for landbruksutvikling, miljøvern og vannressursplanlegging.


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Lastly, I would like to thank my family and friends for their love and encouragement. My friends Hanan Tag El Sir El Safi and Dr. Ahmed Elmokasfi kindly supported me.

My late mother, Suad Khalifa Khogali raised me with a love for challenges and gave me the self confidence to always try and push open new doors. My late father, advocate El Haj El Tahir Ahmed whom I lost at a tender age, continued to enlighten my life through the legacy which he left behind for his love for knowledge and education. My father-in-law Prof. R.S. O’Fahey has tirelessly discussed and outlined the historical and political dimensions of my research through his profound understanding of my home country Sudan. My brothers Khalid and Ayman and my sisters Siza, Dr. Yasmeen, Dr. Safinaz and Sara made my life happier through their respect and understanding. My husband Dominic O’Fahey faithfully supports me in all my pursuits. My children Bushra, Suad and Dalia are a source of inspiration for me and from whom I draw strength to continue working hard.

Thank you all.

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University of Oslo

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Main defence papers:-


In paper I, I was responsible for collecting the field data, remote sensing analyses, writing the draft and final versions of the paper. In paper II, the remote sensing and water balance calculations were made jointly by Wenzhong Wang and I. I was responsible for plotting the results, writing the draft as well as the final version of the paper. In paper III, I was responsible for downloading the data, analysis, draft and final paper-writing.

Additional reference papers:-


### Glossary and abbreviations

<table>
<thead>
<tr>
<th>Abbreviation</th>
<th>Meaning</th>
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<tbody>
<tr>
<td>a.s.l</td>
<td>Above sea level</td>
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<tr>
<td>AE</td>
<td>Actual evapotranspiration</td>
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<td>AI</td>
<td>Aridity index</td>
</tr>
<tr>
<td>ASTER</td>
<td>Advanced Spaceborne Thermal Emission and Reflection Radiometer</td>
</tr>
<tr>
<td>BCM</td>
<td>Billion Cubic Meters</td>
</tr>
<tr>
<td>CAMS</td>
<td>Climate Anomaly Monitoring System</td>
</tr>
<tr>
<td>CPA</td>
<td>Comprehensive Peace Agreement</td>
</tr>
<tr>
<td>DEM</td>
<td>Digital elevation model</td>
</tr>
<tr>
<td>ECMWF/ERA-40</td>
<td>The European Centre for Medium-Range Weather Forecasts</td>
</tr>
<tr>
<td>ENSO</td>
<td>El Niño–southern oscillation</td>
</tr>
<tr>
<td>ET</td>
<td>Evapotranspiration</td>
</tr>
<tr>
<td>EVI</td>
<td>Enhanced vegetation index</td>
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<tr>
<td>FAO</td>
<td>Food and agriculture organization</td>
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<tr>
<td>FCC</td>
<td>False colour composites</td>
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<tr>
<td>GCP</td>
<td>Ground control point</td>
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<tr>
<td>GG</td>
<td>Granger and Gray complementary relationship method</td>
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<tr>
<td>GIS</td>
<td>Geographical information systems</td>
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<tr>
<td>GWR</td>
<td>Geographically weighted regression</td>
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<tr>
<td>Haboob</td>
<td>Dust storm</td>
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<tr>
<td>Hashab</td>
<td>The gum arabic tree (Acacia senegal)</td>
</tr>
<tr>
<td>ITCZ</td>
<td>Intertropical Convergence Zone</td>
</tr>
<tr>
<td>Jabal</td>
<td>Arabic word for mountain</td>
</tr>
<tr>
<td>JRA-25</td>
<td>Japanese reanalysis,</td>
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<tr>
<td>Kerib</td>
<td>Bad lands</td>
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Khor  A gully or a seasonal water course
LAI  Leaf area index
MERRA  Modern Era Retrospective-analysis for Research and Applications
MK  Mann Kendall
MLC  Maximum likelihood classifier
MODIS  Moderate Resolution Imaging Spectroradiometer
NADIR  The downward-facing viewing geometry of an orbiting satellite
NASA  National Aeronautics and Space Administration
NCEP/NCAR  National Centers for Environmental Prediction/National Centre for Atmospheric Research
OLS  Ordinary Least Square
PE  Reference evapotranspiration
PREC  Precipitation REConstruc ted
RMSE  Root Mean Square Error
RS  Remote sensing
SEBAL  Surface Energy Balance Algorithm for Land
SM  Soil moisture
SRTM  Shuttle Radar Topography Mission
SST  Sea Surface Temperature
SSTA  Sea Surface Temperature Anomalies
SWIR  Short-wave infrared
SWIR  Short-wave infrared
TIR  Thermal infrared
VI  Vegetation indices
VNIR  Visible and near infrared
VNIR  Visible and near infrared
WB  Thornthwaite water balance model
The spatial and temporal distribution of soil moisture is a critical part for many disciplines including agriculture, forest ecology, hydroclimatology, civil engineering, water resources, and ecosystem modelling. Despite its importance, long-term soil moisture data are rare because many soil water sampling techniques are expensive, labour-intensive and difficult to learn without appropriate training (Hymer et al., 2000). This is particularly true in data-scarce places like Sudan. Sudan is suffering from rainfall deficit and water shortages. Delayed arrival of the south-westerly flow and associated rainfall often impose great loss of human lives and economy and has disastrous consequences. With much of rural Africa already struggling to obtain adequate fresh water supplies, the drier conditions and altered precipitation patterns mean that meeting the water needs of the poorest in African is much harder. Considerable literature (Thornes, 1990; Ayoub, 1998; Nakileza et al., 1999; Symeonakis and Drake, 2004) points to the widespread natural resource degradation especially in sub-Saharan Africa, including Sudan. Soil erosion results in the loss of soil nutrients, particularly carbon and nitrogen (West, 1991; Mokwunye, 1996), due to the restrictions of feedbacks in carbon and nitrogen cycles between plants, atmosphere and soil (Schlesinger et al., 1990). Often the equilibrium between the vegetation and the environment is delicate and easily displaced due to climate change or interference of man, leading to the destruction of the pattern and the establishment of more xerophytic vegetation. Drought implies some form of moisture deficit and moisture plays an important role in understanding climate change. Therefore the currently well-evidenced global warming is expected to alter the hydrological cycle, and previous studies tend to provide more observed evidences for this viewpoint (e.g., Beniston and Stephenson 2004; Brutsaert 2006; Zhang et al. 2008). Several observational studies, using meteorological data, point to the climate change in Sudan and these studies confirm that the temperature is rising and rainfall has declined in the past decades which might accelerate environmental degradation and desertification (e.g., Alvi, 1994; Janowiak, 1988; Nicholson et al., 2000; Xu et al., 2010; Zhang et al., 2012).

The vegetation-soil moisture relationships describe the various vegetation patterns which occur in Sudan (Wickens and Collier, 1971). These patterns occur in a variety of soils developed from different parent materials. Chemical differences between the vegetated and non-vegetated areas are not significant and the controlling factor is soil moisture. The soils under the vegetation are more permeable to water than the soils in the intervening non-vegetated areas. The infiltration and retention of moisture is determined by the soil type and the character and arrangement of the soil horizons is of prime importance. The retention of a loose sandy surface is very important since it promotes the infiltration of water to the deeper horizons and later, when the surface dries out during drought, it acts as mulch and prevents water loss by reducing capillary movement.
Rainfall is the most important water resource in Sudan and rain-fed agriculture produces food for 90% of the population. Comprehensive analysis and reviews of rainfall trends and variability in Africa, including the Sahel region and Sudan had been carried out by Hulme and his co-authors (e.g., Trilsbach and Hulme, 1984; Hulme, 1987; Hulme and Tosdevin, 1989; Hulme, 1990; Walsh et al., 1988). Trilsbach and Hulme (1984) examined rainfall changes and their physical and human implications in the critical desertification zone in the North and concluded that heavy falls of rain ( > 40 mm) are more likely to decline than medium (> 10 mm) and light (< 10 mm) falls. Walsh et al. (1988) reported that “rainfall decline in semi-arid Sudan since 1965 has continued and intensified in the 1980s, with 1984 the driest year on record and all annual rainfalls from 1980 to 1987 well below the long-term mean”. Hulme (1990) reported that rainfall depletion has been most severe in semi-arid central Sudan between 1921-50 and 1956-85. Similar results are also reported by Eltahir (1992) and Zhang et al. (2012). The length of the wet season has contracted, and rainfall zones have migrated southwards. A reduction in the frequency of rain events rather than a reduced rainfall yield per rain event was found to be the main reason for this depletion. Increasingly the tendency has been for studies examining the causes of droughts, to look at global scales through the notion of teleconnections. The precipitation anomaly of Sudan has been related to Sea Surface Temperature Anomalies (SSTAs) in the Gulf of Guinea (Lamb, 1978a, b). Palmer (1986) has pointed out that the tropical Indian Ocean Sea surface temperature (SST) has a strong influence on the Sahel rainfall. Camberlin (1995) found that the dry and wet conditions over the Sahel were usually associated with warm conditions in the tropical Indian Ocean. The driest years in Sudan were associated with warm El Niño–southern oscillation (ENSO) and Indian Ocean SST conditions (Osman and Shamseldin, 2002). It would seem reasonable that the current drought is a manifestation of an interaction of two or more mechanisms (Cook and Vizy, 2006). Nicholson (1986) pointed out that the variations in the Sahel rainfall are generally related to the changes in the intensity of the rainy season rather than to its onset or length as the Inter-tropical Convergence Zone (ITCZ) hypothesis would require. Vertically integrated moisture flux and its convergence/divergence are closely related to precipitation. Long et al. (2000) aimed to advance the understanding of these causes by examining rainfall, horizontal moisture transport, and vertical motion and how they differ regionally and seasonally. Zhang et al. (2012) analyzed variations in precipitation and the whole layer of moisture fluxes during 1948-2005 in Sudan with the aim of exploring changes in precipitation and associated atmospheric circulation for the occurrence of precipitation transition in Sudan. The annual average precipitation varies greatly in Sudan from almost nil in the north to about 1500 mm in the extreme Southwest. Precipitation decreases almost in all months in the rainy season and significantly decreasing trends can be found during the rainy season and annually. The whole layer of moisture content has significantly decreased in central Sudan in summer since the late 1960s prompting the decline of precipitation in central Sudan. Abrupt changes in monthly and annual precipitation have occurred since the late 1960s. The moisture flux over Sudan tends to decrease after these abrupt changes. The changes in moisture flux are the main reasons for the spatial and temporal variation of precipitation in the region. Much research work has been done regarding the moisture flux over Africa (e.g., Cadet and Nnoli, 1987; Fontaine et al., 2003; Osman and Hastenrath, 1969). Hulme and Tosdevin (1989) found
that changes in the dynamics and flow of the tropical easterly jet exert some control over the Sahelian rainfall. Fontaine et al. (2003) analyzed the atmospheric water and moisture fluxes in the West African Monsoon based on NCEP/NCAR and observed rainfall data and found that at more local scales moisture advections and convergences are also significantly associated with the observed Sudan-Sahel rainfall and in wet (dry) situations, with a clear dominance of westerly (easterly) anomalies in the moisture flux south of 15°N.

Temperature studies by Jones and Lindesay (1993), Elagib and Mansell (2000), and Xu et al (2010) examined long-term changes in an attempt to better understand Sudan’s regional responses to global warming. The annual and seasonal maximum temperatures are increasing significantly. The increasing magnitude of the maximum temperature in the rainy season is larger than that in the dry season. By comparison, the minimum temperature in the rainy season is increasing but the rate of the increase is smaller than that of the maximum temperature in the same season while it is decreasing in the dry seasons. Consequently, the difference between annual maximum and minimum temperature is increasing in all the seasons. There is also a decreasing trend in relative humidity, particularly after mid-1960s, which is in agreement with below-mentioned precipitation changes. The net solar radiation in the region is significantly increasing in all seasons and stations, which corresponds well with the changing properties of the maximum temperature. There is a consistent 1-year period variation within temperature, humidity and net radiation. The climate regime variations of Sudan, to a larger degree, are controlled by global climate signal.

Much water from the reservoirs in Sudan is lost by evaporation and although the drainage area is large, evaporation takes most of the water from the rivers in this region and thus, the discharge of the rivers is small. Reliable evapotranspiration estimates are needed in a wide range of problems, in hydrology, forestry, land management, water resources planning, and irrigation management. For management purposes, actual evapotranspiration has a direct impact on crop yield in rain-fed agriculture in large regions, such as Sudan. Water management in river basins, based on evapotranspiration, has become a developing trend in arid and semi-arid areas (World Bank, 2005; Gao et al., 2012). Compared to traditional management based on water supply and demand, evapotranspiration-based management is more efficient because the utilization of water resources can be managed more efficiently through the reduction of evapotranspiration, hence reducing the overall regional water consumption. Sudan is characterized by huge actual evapotranspiration especially from the vast wetlands in Southern Sudan known as the Sudd, Bahr el Ghazal and the Sobat sub-basins. The evaporation from the Sudd alone is estimated to be more than 50% of the Nile inflow into the north Sudan, i.e. about 28 Gm³/yr out of the 49 Gm³/yr (Sutcliffe and Parks, 1999). The whole river inflow of the Bahr el Ghazal Basin (12 Gm³/yr) is evaporated before reaching the Nile. Therefore planners and engineers suggested methods to save water by reducing the evaporation (losses) from the wetlands and to carry more water to the rapidly expanding population living in the downstream areas in North Sudan and Egypt. Most of the past studies to estimate evaporation in Sudan rely on the computation of evaporation using meteorological ground station data under the basic assumption that the area is wet throughout the year and moisture is not limiting evaporation rates (Sutcliffe and Parks, 1999; and Chan et al., 1980).
Other experiments were made to estimate evaporation from papyrus grown in water tanks by Butcher (1938), however their results were rejected in the subsequent hydrological studies for being too low (1533 mm/yr). More recently actual evapotranspiration was estimated using remote sensing methods (Bashir et al. 2007a; 2007b; 2008; Mohamed et al. 2004; and Abdelhadi et al. 2000).

1.2. Motivation and aims

The above section portrays the scientific background and states the problems associated with soil moisture and their implications in the Sudan. This thesis provides a state of the art, holistic analysis of this valuable resource. There are at best only a handful of previous studies that have looked at soil moisture in Sudan in the thorough way that is presented in this thesis. The thesis tries to answer important questions such as; what are the changing properties of the major climate variables in Sudan as revealed from the historical records? What could be the implications of these changing properties of climate variables on soil moisture and regional water resource management? The answers of these scientific questions are of great importance for better understanding the implications of land-use change on soil degradation in the region.

Specifically, the study objectives are:

- To improve our understanding of the erosion process in Sudan in terms of natural factors and to identify the erosion areas and estimate their changes (Paper I).
- To provide a suitable way for estimation of regional evapotranspiration (Paper II).
- To better understand the trends of rainfall, temperature and soil moisture and the relationship between these major climatic variables and subsequently identify the soil moisture’s spatial and temporal variability especially the dry and wet variations in terms of regeneration demand (Paper III).

To achieve these objectives, a combination of remote sensing, statistical and geo-statistical and modelling methods have been used. The erosion study has benefited from fusion techniques between remote sensing data and geographical information systems (GIS). Regional actual evapotranspiration was best estimated through comparisons of the energy-based remote sensing method, the modified Thornthwaite water balance method (WB), and the complementary relationship method (GG). The Mann Kendall analysis was used to explore the trends of the major climate variables, while the geographically weighted regression (GWR) investigated the relationships of these variables. The aridity index (AI) was used to understand the dry and wet variations in terms of regeneration demand.

1.3. Description of the thesis work

This study demonstrates the spatial and temporal variability of soil moisture and its impacts on soil degradation and vegetation regeneration in Sudan. These impacts are particularly
important in seasonal wet and dry climates. The causes of this variability were studied in terms of regional changes of major climatic variables namely temperature, rainfall and evapotranspiration.

Soil moisture plays a key role in the transfer of energy and mass between land surfaces and the atmosphere, rivers, and aquifers (Hymer et al., 2000). In arid regions, although other factors, especially nutrient availability, influence the behaviour of the ecosystem, soil water availability is recognised as the controlling resource in its organization and function. It is commonly accepted that if water is limited, it becomes the key resource limiting plant growth. It is also one of the main factors that cause plant heterogeneity (Canto’na et al., 2004). However, soil moisture in arid areas is far from being a single homogenous resource; it is highly diversified in several dimensions and it is especially this dimensionality that arid vegetation is adapted to (Krzywinski and Pierce, 2001). Changing conditions of soil moisture have direct impact on soil degradation. In this study soil erosion (Paper I) was first seen as one of the implications of soil degradation. Erosion can be a result of anthropogenic disturbance such as overgrazing, soil crust disturbance, and climatic changes such as precipitation increases or it can happen as a natural process. The important natural factors controlling erosion are, among others, rainfall regime, vegetation cover, terrain, slope, aspect and river flow network (Linsley et al., 1982). Fluctuating rainfall amounts and intensities have significant impacts on soil erosion rates. Where rainfall intensity increases, erosion and runoff increase at an even greater rate. On the other hand, decreasing annual rainfall triggers system feedbacks related to the decreased biomass production that lead to greater susceptibility of the soil to erode (Nearing et al., 2004). Because of the important role of direct rain drop impact, vegetation provides significant protection against erosion by absorbing the energy of the falling drops and generally reducing the drop sizes, which reach the ground. Vegetation may also provide mechanical protection to the soil against soil erosion via the root system. In addition, an adequate vegetation cover generally improves infiltration through the addition of organic matter to the soil. Arid and semi-arid regions are particularly susceptible to soil erosion due to their low plant cover (Bull, 1981). Rates of erosion are greater on steep slopes than on gentle slopes. The steeper the slope the more effective is splash erosion in moving the soil down slope. Overland flow velocities are also greater on steep slopes, and mass movements are more likely to occur in steep terrain. The length of slope is also important. The shorter the slope length, the sooner the eroded material reaches the stream, but this is offset by the fact that overland-flow discharge and velocity increases with length of slope. The river flow network is a representation of the flow accumulation, or the size of the region over which water from rainfall can be aggregated, also known as contributing area. These networks are multiple-branching systems, beginning with tiny rivulets flowing downhill during rainstorms that join into rills and gullies and eventually into creeks and streams. In their headwater regions, river networks are primarily erosional. They acquire soil and weathered rock debris from hill slopes and valley walls. Any abrupt change in any part of the system will propagate uphill across the landscape as well as downstream through these drainage networks (Ritter et al., 2002). As specific catchment area and slope steepness increase, the amount of water contributed by upslope areas and the velocity of water flow
increase, causing stream power and potential erosion to increase (Moore et al., 1988). The influence of these natural factors on soil erosion was one of the concerns of Paper I.

Since soil erosion has a profound impact on the water balance, specifically, on evapotranspiration rates, Paper II was devoted for the estimation of actual evapotranspiration (AE). Actual evapotranspiration describes all the processes by which liquid water at or near the land surface becomes atmospheric water vapour under natural conditions (e.g. Morton, 1983). Globally speaking, AE is the second largest component in the water balance equation, and it is the only term that appears in both the land surface water balance equation and the land surface energy balance equation (Xu and Chen 2005; Romanoa and Giudici 2009). Evapotranspiration (ET) is commonly defined as the loss of water to the atmosphere by the combined processes of evaporation (from bare soil) and transpiration (from plant tissues). Evaporation from the soil surface constitutes a large fraction of the total water loss. Transpiration through crops is regarded as beneficial, but evaporation from bare soils in irrigated fields with a partial canopy cover is considered detrimental (Mehmet et al. 2008). Regional evapotranspiration is perhaps the most difficult term of all the components of the hydrologic cycle to quantify due to the lack of observation data and due to the complex interactions amongst the components of the land–plant–atmosphere system. Its observation needs a lysimeter which is too costly for developing countries. Traditionally, actual evapotranspiration has been estimated as the residual of precipitation and runoff in a long term catchment water balance equation, $E = P - Q$ (e.g. Bosch and Hewlett 1982; Xu and Singh 2004). This method is useful to estimate long-term average annual evapotranspiration (Zhang et al. 2001; 2004) and it is used as a reference to validate other methods. However, this method cannot provide inter-annual, seasonal or shorter term evapotranspiration values for practical use. For these other practical uses, there are other estimation methods. These include the Budyko type evaporation curve as a function of the aridity index in a catchment (Budyko 1974): Different empirical equations of these types exist to describe this relationship (e.g. Schreiber 1904; Ol’dekop 1911; Pike 1964). This type of method uses only annual precipitation and potential evapotranspiration, and is therefore widely used for estimation of annual evapotranspiration (Dooge 1992). Different versions of the Thornthwaite water balance model (Thornthwaite and Mather 1955) have been widely used to calculate actual evapotranspiration at each of the rainfall stations (Xu and Chen 2005; Gao et al. 2007). Several models have been developed based on the Complementary relationship concept (Bouchet 1963; Xu and Li 2003; Xu and Singh 2005). Paradoxically, both advantages and disadvantages of the Complementary relationship method stem from the same fact that it uses only the readily available meteorological data and bypasses the need for soil moisture data which are normally not available and difficult to obtain. Other studies have also reported that its accuracy decreases in water limited environments of arid areas (Xu and Singh 2005; Yang et al. 2008). The method that is developed by the Food and agriculture organization (FAO) based on the Penman-Monteith method (Allen et al. 1998) considers aerodynamic resistance and surface resistance. The drawback associated with this method, however, is that aerodynamic resistance and surface resistance data are not readily available in practice. The remote sensing method (e.g. Courault et al. 2005) has the advantages of covering large areas
and the satellite data are easily obtainable. However, the main drawbacks of the remote sensing method are: the presence of the clouds that contaminate the satellite images; the time cost for calculation of long-term time series; the need for verification of the results by using ground truth data. Due to the complexity of the evapotranspiration process and the lack of observation data, there is no universal method that can give acceptable results everywhere. Each method has its own advantages and limitations. Comparing and contrasting different methods of estimations is expected to result in more accurate assessment of regional evapotranspiration. Paper II compared and contrasted one remote sensing based method with the modified Thornthwaite water balance model and the complementary relationship Granger and Gray (GG) model.

The seasonal differences between wet and dry climates and the spatial differences in actual evapotranspiration prompted a thorough look at the soil moisture variability, its causation in terms of temperature and rainfall and subsequent impact on vegetation regeneration (Paper III).

Soil moisture studies are limited by lack of long-term data due to the amount of resources; both time and money required collecting such data. For example gravimetric measurements of soil moisture are simple and accurate, but are destructive and require at least 24 hours of post processing. Traditional time domain reflectometer sensors yield accurate measurements with calibration, but are expensive and take spatially discrete measurements. Neutron probes are non-destructive and can sample over great depths, but are expensive and potentially hazardous without appropriate training. Fibreglass electrical resistance sensors connected to automated data loggers can record temporally continuous data with little maintenance over long time periods. Like many soil water sensors, they require calibration to convert a measured signal to volumetric water content, something that proved to be difficult (Seyfried, 1993). Alternatively, the National Centers for Environmental Prediction/National Centre for Atmospheric Research, NCEP/NCAR, reanalysis, provides global “soil wetness” values for NCEP R1 (1948 to the present day) and NCEP R2 (1979-present) (Kalnay et al., 1996 and Kistler et al., 2001). The European Centre for Medium-Range Weather Forecasts, ECMWF/ERA-40 provides volumetric soil water in four soil layers (1957-2002) (Simmons and Gibson 2000). The Japanese reanalysis, JRA-25 includes a dimensionless global data product called “soil wetness” (1979-2004) (Onogi, et al., 2007) and the Modern Era Retrospective-analysis for Research and Applications, MERRA, (1979-Present) (Bosilovich 2008). In this study NCEP/NCAR R1 soil moisture and temperature estimates are used because it provides a longer time period compared to the other global reanalysis and because of that it has received much more use and attention than others. Also there are only minor differences that are found between R1 and R2 (Kanatmitsu et al., 2002). Because of the limited availability of observed data, rainfall data for Sudan was downloaded from the global monthly precipitation data (PREC) created by Chen et al. (2002) instead of NCEP/NCAR precipitation data. This is because previous studies have shown that the NCEP/NCAR reanalysis precipitation exhibits some deficiencies and this field does not compare well with observations as other reanalysis fields do, such as soil moisture and temperature that are assimilated directly into the model (Poccard et al., 2000; Janowiak et al., 1998). The PREC
analyses are derived from gauge observations from over 17000 stations collected in the Global Historical Climatology Network (GHCN), version2, and the Climate Anomaly Monitoring System (CAMS) datasets. Shi et al. (2002a, 2002b, 2004) analysed the global land precipitation dataset (PREC/L) and found that this dataset was accurate enough for describing the change of large scale precipitation, and there was a good relationship between the global precipitation database (PREC) and observed Western Africa monsoon precipitation during 1948-2001. PREC data has, in general, a good agreement with observed data in Sudan (Zhang et al., 2012). In this study NCEP/NCAR and PERC data (Paper III) are used to study the general trend and to understand the SM, temperature and rainfall relations. Because the aim here is the general trend and variability and not a detailed or specific study, such as a water balance study, these data are thought to be sufficient. However, for the estimation of actual evapotranspiration, Paper II, daily observed data were used.

The study has focused on the top soil layer because it is this layer that is more important for the regeneration of perennials. The hypothesis is that plants require a minimum amount of soil water at the top layer before regeneration of perennials occurs. The hypothesis focuses on the regeneration of perennials because it is the main constraint in permanently controlling desertification (Li et al., 2004). Additionally, the top layer is more relative to remote sensing data that were used in the study. Remotely sensed satellite observations are only able to measure soil moisture within 1-2 cm of the surface but can retrieve information that well represents the top 10 cm layer (Vinnikov et al., 1999).
2. Study Area

2.1. About Sudan

The choice of the study area in each paper depended on the study objectives, and the availability of data and information. The study area in Paper I covers the area bounded by latitudes 11° and 16° N and longitudes 33° and 35° E in Eastern Sudan, traversed by the Blue Nile. Paper II, also in Eastern Sudan, included three meteorological stations: Abu Naama, Damazine and Gedarif that lay between latitudes 11.8° and 15.5° N and longitudes 32.2° and 34.5° E. In Eastern Sudan (Paper I & II), observed data indicated that the mean annual precipitation in 2006 was 900 mm, of which around 90% was collected during the rainy season (April–October). Paper III covered the whole country.

Figure 1-1 Map of Former Sudan showing the spatial distribution of average annual soil moisture (a) and rainfall (b) revealing four distinct climatic regions: North-Arid; Centre-Semi arid; South-Humid and South-Tropical

(This figure is re-produced from Paper III)
**General:** The Former Sudan (Fig. 1-1), was the largest country in Africa with an area of about 2.5 million km$^2$. It extends over 17 lines of latitude from the Sahara region in North Africa to the Equator and 19 longitudes from the Congo basin in central Africa to the west, to the Red Sea in the East. The latest 5th Sudan Census of 2008 estimated the country’s population at about 39.5 millions (United Nations Population Fund [UNFPA], 2008). The climate ranges from arid in the north to tropical wet-and-dry in the far southwest. About two-thirds of Sudan lies in the dry and semi-dry region. There are major contrasts in rainfall (P) and potential evapotranspiration (PET), thus producing distinct climatic zones ranging within arid (North); semi arid (Centre); humid (South) and tropical (South), as classified using P/PET ratio (United nations Environmental Programme [UNEP], 1992; Hare and Ogallo, 1993). Sudan is drained mainly by the Nile River and its two main tributaries, the Blue and the White Nile. The Nile River is the longest river in the world, flowing for 6,737 km from its farthest headwaters in central Africa to the Mediterranean and is the lifeline for Sudan.

**Rainfall:** the most significant climatic variables are rainfall and the length of the rainy season. From January to March, the country is under the influence of the dry northeasterly winds, and during this period there is almost no rainfall countrywide except for a small area in north eastern Sudan where the winds pass over the Mediterranean bringing occasional light rains. By early April, the rainy season starts from southern Sudan as the moist south westerlies reach the region, and by August the southwesterly flows extend to the northern limits. The dry northeasterlies begin to strengthen in September and to push south and cover the entire country by the end of December. Rainfall is negligible in the North, whereas equatorial climatic conditions prevail in south with annual rainfall of more than 1015 mm yr$^{-1}$. Rainfall amounts decrease from the south to the north and the dry season lasts between three months in the humid tropical south and nine months in Khartoum with the hottest months being July and August. Since the 1960’s, rainfall has declined considerably.

There are two areas of interest in Sudan’s climate (El Tahir, 1989). These are the Bahr El Ghazal Basin in the southwestern corner of the country and the coastal area that forms a small strip along the Red Sea in the north-eastern part. Bahr El Ghazal Basin is characterised by high rainfall levels, with a rainy season of eight months. Two processes dominate the hydrology of Bahr El Ghazal basin, namely rainfall and evapotranspiration, while all the other processes are negligible. The Red Sea area lies in North-Eastern Sudan. The prevalent climatic conditions are characterised by low precipitation and high evapotranspiration. The mountains and the sea distinguish the area from the rest of the country and they influence the micro-climate within the region. Two contrasting rainfall regimes exist within this region. The area to the west of the mountain ranges experiences maximum rainfall during the summer (July/August) brought about by the south westerly monsoon winds which originate from the gulf of Guinea, whereas the area to the east receives maximum rainfall during the winter brought about by the north east and south east winds blowing over the Red Sea.

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1This thesis was written before the sessesion of South Sudan in July-09-2011, therefore we refer here to former Sudan meaning both the North and the South.
**Temperature:** Temperatures do not vary greatly with the season at any location compared with many regions of the world. Summer temperatures often exceed 43.3°C in the northern desert zones. Dust storms, known locally as haboob, frequently occur. High temperatures also prevail to the south throughout the central plains region, and the humidity is generally low. In the vicinity of Khartoum, the capital of Sudan, the average annual temperature is about 37.1°C. January is the coolest month (with mean minimum 15°C and mean maximum 30°C). June is the hottest month (with mean minimum 38°C and mean maximum 42°C). In southern Sudan the average annual temperature is about 29.4°C.

**Evapotranspiration:** Similar to rainfall and temperature, annual actual evapotranspiration varies between the different climatic zones. June-September have the highest values while November-February have the lowest values of evapotranspiration. When relating potential evapotranspiration and rainfall values it appears that the area suffers climatically from acute water deficit.

**Soil:** The country's soils can be divided geographically into three categories. These are the sandy soils of the northern and west central areas, the clay soils of the central region, and the laterite soils of the south. Less extensive and widely separated, but of major economic importance, is a fourth group consisting of alluvial soils found along the lower reaches of the White Nile and Blue Nile rivers, along the main Nile to Lake Nubia, in the delta of the Qash River in the Kassala area, and in the Baraka Delta in the area of Tawkar near the Red Sea. Agriculturally, the most important soils are the clays in central Sudan that extend from the East through central to southwestern Sudan. Known as cracking soils because of the practice of allowing them to dry out and crack during the dry months to restore their permeability, they are used in the areas of the Gezira Scheme in central Sudan (the largest user of the Nile waters), the Rahad Project, the New Halfa Scheme, the Suki Scheme, the White Nile and Blue Nile Pumps Schemes, and the Kenana Sugar Scheme for irrigated cultivation. East of the Blue Nile, large areas are used for mechanized rainfed crops. West of the White Nile, these soils are used by traditional cultivators to grow sorghum, sesame, peanuts, and (in the area around the Nuba Mountains) cotton. The southern part of the clay soil zone lies in the broad floodplain of the upper reaches of the White Nile and its tributaries. Subject to heavy rainfall during the rainy season, the floodplain proper is inundated for four to six months, and a large swampy area, the Sudd, is permanently flooded. Adjacent areas are flooded for one or two months. In general this area is poorly suited for crop production, but the grasses it supports during dry periods are used for grazing. The sandy soils in the semiarid areas support vegetation used for grazing. Livestock raising is a major activity, but a significant amount of crop cultivation, mainly of millet, also occurs. Peanuts and sesame are grown as cash crops. These sandy soils are the principal area from which gum arabic is obtained through tapping of Acacia senegal (known locally as hashab). The laterite soils underlie the extensive moist woodlands found in South Sudan. Crop production in these soils is scattered, and the soils, where cultivated, lose fertility relatively quickly; even the richer soils are usually returned to bush fallow within five years.
**Vegetation:** Open acacia communities are found in hill valleys and catchment areas of inland and coastal plains with bushes and trees of Acacia tortilis, a. radiana, a. etbaica, Salvadora persica, and Maerua crassifolia. Annuals and ephemerals include Cenchrus spp, panicum, Euphorbia spp, aristida, etc. Plant cover is less than 40% in the whole country, a percentage that may increase up to 50% due to seasonal growth. It portrays different land use types including agriculture, forests and range areas. Different agriculture forms include small holdings, mechanized and river bank farming. The area exhibits high land cover variability. It includes for example the Ar Roseris power dam and its lake. There are a number of forests for example the Sunut forest reserve and Okalma forest reserve. The forest reserve is a natural forest composed of a mixture of trees, mainly different types of Acacias, Balanites aegyptiaca plus other species. There is a large variation of age groups from young new regeneration to large groups of forest stands. The forests are related to the surrounding societies providing diversified benefits from the trees and the land. Regeneration and forest development factors are evident. Mountains include the Red Sea Hills in the North East, the Jabal Marra in the west, the Ingasana Hills is the south East, and the Nuba Mountains in the South West. The mountains constitute a source of sheet floods. Other areas are abandoned fallow land formed following abandonment of agricultural cropping. There are also the seasonal Khors (a gully or a seasonal water course is locally known as khor), running in different directions. The khors are characterised by the presence of degraded areas, natural regeneration, human activities such as agriculture, pastoralism and small settlements, as well as human intervention to reclaim the vegetation.

**Topography and the Khor System:** The Khor system can be observed on a wide range of spatial scales, and each unit offers different environmental conditions for plant growth. Because of these broad scale geomorphological and finer scale topographic variations, there is an uneven distribution of water in the landscape, and hence variation in soil moisture. The moisture gradient associated with topographical variations has a strong influence on plant distribution. The transect that represents the different units in the landscape may also be viewed as a gradient, due primarily to variations in soil depth and soil moisture, which co-vary from the desert pavements to khors. The distribution of the available water resources is governed by a system of these khors. It is along the sides and on the beds of these khors that trees are found, agriculture is practiced, and where hand dug wells that provide the main water supplies are found. Gully erosion stripes off the fertile clay soils from the degradational clay plain forming bad-lands known locally as “Kerib” (Mirghani, 2007). Although erosion in the centre of the gully is visually apparent, its effects are not always detectable in terms of changes in soil quality (Ward et al., 2001). This indicates how geomorphologically-apparent desertification (Nir and Klein, 1974; Rozin and Schick, 1996) and changes in soil nutrient content are not necessarily congruent. Nonetheless, a decline in soil nutrients is recorded as some of the most important soil variables, and thus is likely to significantly impact plant growth. Although the major erosion usually occurs in the centre of the gully in a strip that is only 5–30m wide, most plant biomass and species diversity are well concentrated there (Ward and Olsvig-Whittaker, 1993). Moreover, the concentration of the water current in the central erosion gully necessarily reduces the water flow to the adjoining
sides of the valley during floods. As a consequence of this reduction in water availability, leaching of salts is reduced (Shalhevet and Bernstein, 1968; Dan et al., 1973; Dan and Yaalon, 1982; Dan and Koyumdjisky, 1987) and soil salinity increases on the sides of the valley. Thus, even in the soil that remains un-eroded, soil quality declines over time.

2.2. Geopolitical and economic setting

The White Nile and its tributaries Sobat River, Bahr el Arab, Bahr el Ghazal, Bahr el Zeraf, Lol, Yei, Jur, Tonj, and Naam rivers, originate from Lake Victoria and Lake Albert in Uganda and also from Ethiopia and it contributes about 28% of the total flow of the Nile River. The Blue Nile and its tributaries, including the Rahad and Dinder rivers, rise in the Ethiopian highlands and it contributes about 59% of the total flow of the Nile River. Upon their confluence at Khartoum, the White Nile and the Blue Nile form the Nile River. The Nile River is joined after that, still in Northern Sudan, by the Atbara River which also originates in the Ethiopian highlands and contributes about 13% of the total flow of the Nile River. The Atbara River is the last tributary to join the Nile which thereafter flows through Northern Sudan and Egypt before emptying into the Mediterranean Sea (Collins, 2002). Despite the high contribution of the Blue Nile, its peak flow is largely seasonal, concentrated mostly in the months of June through to September and it is laden with silt that it carries over from the Ethiopian highlands. On the other hand, the relatively smaller contribution of the White Nile is mostly steady throughout the year, and, it is almost silt-free thus providing for the critical water needs of Sudan and Egypt during the low flow period of the Blue Nile. In addition to the Nile water Sudan has vast groundwater resources. Indeed, a number of groundwater basins such as the Upper Nile Basin fall across the borders between the North and the South, largely fed and replenished by the White Nile and its tributaries. The Nubian sandstone aquifer is shared by Sudan, Libya, and Chad. However, technical knowledge and data about these aquifers are, at best, quite limited, and therefore this resource remains largely untapped.

The waters of the Nile are governed mainly by two agreements. The first is the 1959 Nile Waters Agreement which allocated 18.5 Billion Cubic Meters (BCM) to Sudan and 55.5 BCM to Egypt (Salman, 2010). Sudan’s average use has so far ranged between 14 and 16 BCM annually. This agreement created a sense of animosity by some of the other eight Nile riparians countries namely Ethiopia (where 86% of the Nile waters come from), Eritrea, Uganda, Kenya, Tanzania, Democratic Republic of Congo, Rwanda, and Burundi of their right to an equitable and reasonable utilization of the Nile waters. The second agreement is the “Wealth Sharing” Agreement which is part of the 2005 Comprehensive Peace Agreement (CPA) between North and South Sudan². Although this agreement provided detailed

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² The deal signed on 9 January 2005, between the North and the South known as Comprehensive Peace Agreement (CPA) brought temporary peace to Sudan ending Africa’s longest running civil war. The deal consisted of several protocols and agreements that governed the relationship between the North and the South during a six-year transitional period which ended with the referendum on self-determination for the South (Sudan Open Archive, 2005) in January 2011, and which was followed by the official birth of the new Republic of South Sudan on 9-July-2011.
provisions on sharing and management of land and natural resources in areas such as, land sharing and management, sharing of the oil revenues, monetary policy, banking, currency, borrowing, etc, it sadly omitted any but brief mention of water resources. Moreover, the agreement did not make an explicit mention of the groundwaters shared between the North and the South of the Sudan.

The hydro-politics of the Nile Basin are facing yet again a new challenge. Following the declaration of a new country, the Republic of South Sudan in July-09-2011, the total number of the Nile Basin riparian countries rose to eleven. The new country has developmental plans that involve using water resources for the generation of hydro-power, irrigation, as well as for domestic, livestock and industrial uses. The water will be provided from either surface and/or underground waters shared with one or more of these countries.

At the backdrop of this complex geopolitical situation, this thesis is an authentic addition to key technical knowledge about this important but often ignored resource.
3. Methods

In order to achieve the objectives that are stated above, a number of methods have been used, developed and/or modified. The methods used are grouped into three categories: the remote sensing methods; the statistical and geostatistical methods and the modelling methods. The details of the methods are outlined below according to this classification.

3.1. Remote sensing methods

- Erosion model

The flow diagram in Fig. 4-1 demonstrates the method used in developing the erosion area model (Paper I). Briefly speaking, the study area was divided into two scales. The first, a 60×60 km (approx. 3600 km²) scale, was studied in detail with the aid of the fine resolution of two ASTER images dated 30th March and 18th December 2006. After verifying the ASTER results, two MODIS images dated 22nd March and 19th December 2006 were used to scale up the results to the second 180×180 km (approx. 32 400 km²) scale. The aim was to detect the bi-temporal changes in eroded area that have resulted from the seasonal rains of 2006.

An initial preprocessing of the raw data was undertaken. The pre-processing of ASTER involved a) generation of two photogrammetric DEMs from the two ASTER images, b) comparison of the accuracy of the said ASTER DEMs with SRTM, c) orthorectification of the images, and d) layer stacking.

DEM generation: Because SRTM has significant voids especially in high mountains that result from radar shadow, foreshortening, layover and insufficient interferometric coherence, DEMs were generated from ASTER. DEM generation from satellite imagery uses

Figure 3-1 Erosion model. ASTER (left) and MODIS (right) processing steps to estimate the change in area of eroded land between March and December 2006.

(This figure is re-produced from Paper I)
photogrammetric principles. Toutin (2001, 2002), and Toutin and Cheng (2001, 2002) outlined the main digital processing steps for DEM generation from ASTER within the PCI Geomatica software. Briefly, the first three (NADIR) channels 123N and the back looking channel 3B of each ASTER were read as two separate .pix files in PCI Geomatica’s orthoengine using Toutin’s low resolution model. The scenes were reprojected to UTM WGS1984 Zone 36N projection. The ground control points (GCPs) necessary for estimating the model parameters were collected from a hillshade image of SRTM generated in ArcGIS 9.3. Due to the lack of accurate topographic maps for the study area, the hillshade of SRTM was used. The hillshade SRTM ensured better co-registration between the SRTM and ASTER derived products later in the combination process between the two. In total 29 GCPs are collected for each image. The two images 123N and 3B were tied together using 95 tie points. Using cubic interpolation, a 60m resolution DEM was extracted from each ASTER image, named accordingly March DEM and December DEM.

The qualities of ASTER and SRTM DEMs were compared and the most accurate of them was used further for orthoprojection. Because the ASTER DEMs cannot be tested against an existing reference DEM, they are tested through the overlay of different orthoimages taken from different sensor positions (also know as multi-incidence angle image). Quantitatively, the generated DEMs were tested by producing two orthoimages from each ASTER, one for channels 123N and another for channel 3B using the respective generated DEM. The two orthoimages were then overlaid to test if they overlap perfectly well pixel-by-pixel using animation flickering techniques. Quantitatively, the error of ASTER DEMs, is also estimated from their contour lines when compared to the contour lines of SRTM. The tested contours represented rugged high-mountain conditions with elevations of up to 1500 m a.s.l. with steep rock walls and deep shadows that are without contrast. Therefore, the test area is considered to represent a worst case for DEM generation from ASTER data.

After comparisons the SRTM DEM was found to be more accurate than the ASTER DEMs, and therefore it was used for orthoprojection. Orthoprojection is a mandatory pre-processing step necessary to prevent strong topographically induced distortions between the images in the rugged study area which is characterised by variable elevation ranging approximately from 10 m to 1500 m.

After orthoprojecting ASTER images, the channels that best described the process of soil erosion were selected. Considering the spatial width of gullies, ASTER’s VNIR and SWIR channels at 15 m spatial resolution were analysed. SWIR has 30 m resolution but was resampled to 15m. Additional layers that represent some of the most important natural factors causing erosion were added. These layers were elevation, slope, aspect, and river flow network. Therefore the final data set used for classification consisted of 13 layers stacked together. These included: VNIR channels 1, 2 and 3, SWIR channels 4, 5, 6, 7, 8 and 9, SRTM DEM elevation, slope, aspect, and river flow network. The last three layers; slope, aspect and river flow network were all calculated from SRTM in ArcGIS 9.3. Both slope and aspect were calculated from SRTM using a finite difference method that used eight neighbours in ArcGIS. The two were then resampled to 15 m before adding them to the layer
The river flow network was estimated using the hydrologic analysis package in ArcGIS 9.3. When working with raster DEMs and computing slopes between grid cells, the ratio of the vertical and horizontal resolutions determines the minimum non-zero slope that can be resolved. In this study, the vertical resolution of the DEM was 1 m and the grid size was 15 m. Therefore the resolvable slope was calculated as $1/15=0.07$. This value means that slopes on hillsides can be computed with a relatively small error. However, slopes in channels were often much smaller than this value. As a consequence, these areas appeared horizontal in the DEM. Therefore in order to avoid this problem, ArcGIS 9.3 uses an eight direction (D8) flow model. The direction of flow was determined by finding the direction of steepest descent, or maximum drop, from each cell. This was calculated as maximum drop = change in z-value/distance. The distance was calculated between cell centres. There are eight valid output directions relating to the eight adjacent cells into which flow could travel. One challenge arises if all neighbours are higher than the processing cell. In such a case the processing cell is called a sink and has an undefined flow direction because any water that flows into a sink cannot flow out. To obtain an accurate representation of flow direction across a surface, the sinks should be filled. The minimum elevation value surrounding the sink identifies the height necessary to fill the sink so the water can pass through the cell. A digital elevation model free of sinks is called depressionless DEM. Using the depressionless DEM as an input to the flow direction process, the direction in which water would flow out of each cell was determined. After determining the flow direction, then the flow accumulation was determined. Afterwards, a stream network was created by applying a threshold value to select cells with high accumulated flow in order delineate the stream network. More details on the technique of deriving flow direction from a DEM and on how to create river network in ArcGIS can be found in Jenson and Domingue (1988) and in Tang and Liu (2008).

After successful pre-processing the resulting orthorectified images were analysed. The image analysis involved supervised maximum likelihood classification aided by the authors’ knowledge of the area. Each image was trained into four land cover types: Gully, Flat land, Mountain and Water. Water bodies were clear and easy to train considering the fine ASTER resolution for both periods of the year. Mountains were also easily classified with the aid of the DEM. However, the most challenging task was to separate between Gully and Flat land since these two classes are bound to overlap and overlapping training area boundaries reduces the reliability of the training sites. To avoid this kind of overlap, a number of steps were taken. These steps included using MODIS vegetation indices (VI) as auxiliary data to discriminate between training classes. At first MODIS NDVI and EVI signatures are used to discriminate between stable and unstable vegetation. The unstable vegetation was seasonal and grew during the rainy season. That vegetation was flushed away with erosion, indicating that areas where there was unstable vegetation there was also erosion. The stable vegetation on the other hand was there throughout the year hence no erosion occurred. Where NDVI and EVI values are low, this is an indication of limited, unstable vegetation hence higher erosion risk. By contrast high values of NDVI and EVI indicated more stable vegetation. A second means of discriminate between Gullies and Flat land was by superimposing river network layer on top of the image to help guide the classification process. Other measures for ensuring
accurate classification included selecting bands that have better display, in either greyscale or as false colour composites FCC. Additionally, the classes are refined by varying the number of training areas until better accuracy is achieved. Better accuracies were also obtained through training as many areas as possible. Once the training areas are defined, then the signature separability values were studied. Signature separability is the statistical difference between pairs of spectral signatures. It is expressed in terms of Bhattacharryya Distance and Transformed Divergence. These are measures of the separability of a pair of probability distributions. Both Bhattacharryya Distance and Transformed Divergence are shown as real values between zero and 2. Zero indicates complete overlap between the signatures of any two classes while 2 indicates a complete separation between the two classes. The higher the separability value (i.e. more than 1.5) the more accurate the classification accuracy. The training areas are tuned until higher signature separability value of 1.5 or more are achieved. After achieving the best accuracies, the signature statistics report was studied in order to determine which channels of the 13 stacked layers were more significant in delineating erosion gullies.

Post classification steps were:

a) ASTER classification is validated via running an automatic random accuracy assessment analysis. This analysis was designed and implemented by generating a random sample of 300 points and comparing them to the original ASTER image. Each of the 300 samples was assigned to the different classes.

b) The area represented by each of the four classes is calculated using the function Generate Area Report in PCI Geomatica. Areas in the December image are subsequently subtracted from their correspondent areas in the March image in order to calculate the changed area per class.

Once ASTER classification results were accepted, the results were then up-scaled using MODIS. Upscaling was carried out by using the same training areas from ASTER, which were converted to shape files and superimposed on the MODIS products to guide the classification of the latter. In this way the results of the of ASTER classification were generalized to larger areas covering the whole of the Blue Nile region. MODIS images were first georeferenced before undertaking a supervised classification using the same land cover classes and training areas as ASTER. Additionally, the classification was aided by using MODIS NDVI and EVI and river flow network layers as auxiliary data to discriminate between training classes. The outcome of MODIS classification was validated against field data using digital photos with co-ordinates and time taken in January 2007 from different locations. Finally the seasonal change in erosion area for the whole Blue Nile region was estimated.
SEBAL model

SEBAL (Waters et al. 2002) is an instantaneous, i.e., time is constant, energy balance model. It considers the fact that evapotranspiration consumes energy and accordingly evapotranspiration ($AE$) is calculated as a residual of the energy balance (Eq. 1)

$$\lambda AE = R_n - G_s - H_s$$  \hspace{1cm} (1)

Where $\lambda AE$ is the instantaneous latent heat flux representing the energy available for evapotranspiration; $R_n$ is the net radiation; $G_s$ is soil heat flux, i.e., rate of heat storage into the soil and vegetation due to conduction; and $H_s$ is the sensible heat of flux. All the components have the units of (w m$^{-2}$).

In the SEBAL model, the net radiation is computed from spatially variable reflectance and emittance of radiation. The calculation requires spectral radiances in the visible, near infrared and thermal infrared regions of the spectrum to determine the intermediate parameters, such as surface albedo, NDVI and surface temperature. The soil heat flux, $G_s$, is computed as an empirical fraction of the net radiation using surface temperature, surface albedo and NDVI as input variables. The estimation of sensible heat flux, $H_s$, requires an iterative solution until $H_s$ converges to the local non-neutral buoyancy for each pixel. The original MOD-1B level 1 versions 4 and 5 daily calibrated radiance with spatial resolution of 1 km and 36 bands for the year 2006 (available from [http://ladsweb.nascom.nasa.gov/data/search.html](http://ladsweb.nascom.nasa.gov/data/search.html)) was used. Fourteen MODIS images dating from 07-January through 20-December 2006 were used. Each month was represented by minimum one image. MODIS’s NDVI is chlorophyll sensitive and it demonstrates a good dynamic range and sensitivity for monitoring and assessing spatial and temporal variations in vegetation amount and condition (Huete et al. 2002). It is successful as a vegetation measure in that it is sufficiently stable to permit meaningful comparisons of seasonal and inter-annual changes in vegetation growth and activity. In this study, the DEM is used to calculate the land surface albedo, emissivity, surface temperature over the horizontal plain of the image, and the surface roughness. These parameters are needed to solve Eq. 1 above. Together with land surface temperature and albedo, NDVI is used to calculate the leaf area index (LAI). LAI, in turn, is used to calculate Local surface roughness index ($Z_{0m}$) for short grass or crop vegetation. Ultimately the Zero plane displacement height (d) is calculated. [$Z_{0m}$] and [d] values are directly available from NASA’s table of Mapped Monthly Vegetation Data at [http://ldas.gsfc.nasa.gov/nldas/NLDASmapveg.php](http://ldas.gsfc.nasa.gov/nldas/NLDASmapveg.php). Briefly speaking, the remote sensing procedure involved georeferencing and bands selection of MODIS data; mosaicing and subsetting of the different images to fit the study area; preliminary analysis using false colour composites (FCC), instantaneous energy calculation using SEBAL model using Eq. 1 above and daily energy calculation using SEBAL model. The detailed calculation procedure of SEBAL model is available, among others, from Bastiaanssen et al. (1998a; 1998b; 2000; 2002; 2003), Mohamed et al. (2004), and Bashir et al. (2007a; 2007b).

3.2. Statistical and geostatistical methods

- Test of autocorrelation

Before undergoing trend analysis, the serial correlation of the data was studied, because the presence of serial correlation would affect the detection of trends in a series (Serrano et al.,
The analysis for checking serial correlation is conducted by examining the lag-1 serial correlation coefficient (designated by $r_1$) of the time series, which is calculated using the equation (2) by Haan (2002):

$$ r_1 = \frac{1}{n-1} \sum_{i=1}^{n-1} (X_i - \bar{X})(X_{i+1} - \bar{X}) $$

$$ \frac{1}{n-1} \sum_{i=1}^{n-1} (X_i - \bar{X})^2 $$

(2)

Where $n$ is the sample size and 1 is the time lag.

The critical value for a given significance level, $\alpha$ is calculated by:

$$ p / q = -1 \pm z_{\alpha/2} \sqrt{\frac{n-k-1}{n-k}} $$

(4)

Where $p$ and $q$ are the upper and lower limits, respectively. $z$ is the critical value of the standard normal distribution for a given significant level $\alpha$ (5% in this case). If the calculated $r_1$ was not significant at the 5% level, then the trend test was applied to original values of the time series. If, however, the calculated $r_1$ was significant, prior to application of the test, the “de-seasoning” of the time series was obtained as (Partal and Kahya 2006; Luo et al., 2008; Chu et al., 2010):

$$ x_2 - r_1 x_1, x_3 - r_1 x_2, \cdots, x_n - r_1 x_{n-1} $$

(5)

### Trend analysis

The Mann-Kendall trend test (Mann, 1945; Kendall, 1975) is regarded as a powerful tool for exploring trends in hydrological series. It is the most widely applied non-parametric test for detecting a trend in the hydro-meteorological variables with respect to climate change or variability (Abdul Aziz et al., 2006; Chen et al., 2007). Parametric and non-parametric methods have been extensively applied for detection of trends. Parametric tests are more powerful than the non-parametric tests, but there is an implicit assumption of normality of data that is seldom satisfied. Hydro-meteorological time series like SM are often characterized by data that are not normally distributed, and therefore nonparametric tests are considered more robust compared to their parametric counterparts (Hess et al., 2001). The Mann-Kendall test is distribution-free and therefore no assumption regarding the distribution of data is needed. A major advantage of this test is that it allows for missing data and can tolerate outliers. Mann Kendall test is a rank based approach that consists of comparing each value of the time series with the remaining in a sequential order.

The statistic $S$ is the sum of all the counting as given in equation (6) below;

$$ S = \sum_{i=1}^{n-1} \sum_{j=i+1}^{n} \text{Sgn}(X_j - X_i) $$

(6)
Where

\[ S_{\text{gn}} = \begin{cases} 1 & \text{if } (X_j - X_k) > 0 \\ 0 & \text{if } (X_j - X_k) = 0 \\ -1 & \text{if } (X_j - X_k) < 0 \end{cases} \]

(7)

and \( X_j \) and \( X_k \) are the sequential data values; \( n \) is the length of the data set. A positive value of \( S \) indicates an upward trend and a negative value indicates a downward trend. The standardized test statistic \( (Z_{mk}) \) is calculated by

\[
Z_{mk} = \begin{cases} \frac{S - 1}{\sqrt{\text{Var}(S)}} & \text{if } S > 0 \\ \frac{S + 1}{\sqrt{\text{Var}(S)}} & \text{if } S < 0 \\ 0 & \text{if } S = 0 \end{cases}
\]

(8)

Where the value of \( Z_{mk} \) is the Mann-Kendall test statistics that follows standard normal distribution with mean of zero and variance of one. Thus, in a two sided test for trend, the null hypothesis of no trend, \( H_0 \) cannot be rejected if

\[
-Z_{1-\alpha/2} \leq Z_{mk} \leq Z_{1-\alpha/2}
\]

(9)

Where \( \alpha \) is the significance level that indicates the strength of the trend. In the present study, significance level of 5% was applied.

– Linear regression

The relationships between soil moisture and rainfall and temperature inherently vary over space, and can be analyzed using geographically weighted regression (GWR) technique. GWR, a recent refinement of ordinary linear regression models, is a local spatial statistical technique used to analyse spatial non-stationarity when the measurement of relationships between variables differs from location to location. First the relationship was explored using the following equation of the ordinary least square regression (OLS) below:

\[
SM = B_0 + B_1 \text{Temperatur} + B_2 \text{Rainfall} + \varepsilon
\]

(10)

In this model, the dependent variable was mean monthly soil moisture (mm), the two independent variables were mean monthly temperature (°C), and average monthly rainfall (mm). \( B_0, B_1 \) and \( B_2 \) regression coefficients, and \( \varepsilon \) is the residual. This model assumed spatial stationarity in the relationship among the different variables under study.

In comparison to the OLS, the GWR base equation used in the multivariate regression analysis is:

\[
SM (u,v) = B_{0}(u,v) + B_{1}(u,v) \text{Temperatur} + B_{2}(u,v) \text{Rainfall} + \varepsilon (u,v)
\]

(11)
Where \((u,v)\) are the location coordinates of a particular point on a surface.

This method takes into consideration the local variations in rates of change with resulting coefficients calculated by the model that are specific to each location (Brundson et al., 1996). Therefore, it has emerged as one of the more effective methods to account for, and to examine the presence of spatial non-stationarity in relationships (Fotheringham et al., 1998). In particular, the parameter estimates are calculated by using a weighting method in which the contribution of an observational site to the analysis is weighted in accordance with its spatial proximity to the specific location being considered. Thus, the weighting of an observation in the analysis is not constant but a function of its location. Observations close to the centre of the window are weighted more heavily than observations further away. The derived parameters can then be mapped to identify the nature of their variation in space and thus bring out the spatially varying relationships among the different variables. In this method, parameter estimation is highly dependent on the weighting function and kernel used.

NDVI data were used as weight variables in the analysis. The GWR analysis was conducted using an adaptive kernel method. A small bandwidth results in very rapid distance decay, while a large value results in a smoother weighting scheme. The variable bandwidth approach was chosen and optimized by cross-validation minimisation to account for the spatial variation in the size of the different climatic regions, and hence the density of these regions’ centroids.

- **Aridity Index (AI) and irrigation requirements**

To further understand dry and wet variations in terms of regeneration demand, the aridity index is calculated in this study (Paper III). Aridity index, \(AI\), as defined by de Martonne (1926) who used it to study irrigation demands (Zhang et al., 2009) is computed as:

\[
AI = \frac{12 \, P}{T + 10}
\]

Where \(P\) is the monthly precipitation amount; \(T\) is the monthly mean air temperature.

The aridity index aims to identify the months when irrigation is necessary. Generally, irrigation is necessary when \(AI < 20\).

### 3.3. Modelling methods

- **The modified Thornthwaite Water Balance Model**

The Thornthwaite water balance model (Thornthwaite and Mather 1955) is usually applied with monthly or daily rainfall time step. The use of daily values is preferred over monthly values when possible, particularly in dry locations, where precipitation only occurs occasionally except in the rainy season. In such a case monthly totals cannot reflect the dynamics of soil moisture. In this study, the water balance calculations are performed on a daily time step. There are different variants of the Thornthwaite water balance model, and the basic procedure of the model used in this study is similar to that used by Xu and Chen (2005).

The soil moisture storage is computed as:
Where \( W_t \) is the current soil moisture, \( W_{t-1} \) is the soil moisture in the previous time step, \( P_t \) is precipitation, and \( AE_t \) is actual evapotranspiration at time step \( t \).

- If potential evapotranspiration \( PE_t \leq P_t \), then \( AE_t = PE_t \). \( W_t \) is first estimated with surplus (runoff and/or recharge) = 0.

  When \( W_t > FC \) (field capacity), surplus = \( W_t - FC \), and \( W_t \) is set to FC;

  When \( W_t \leq FC \), Surplus = 0.

- If \( PE_t > P_t \), soil water will be depleted to compensate for the supply. At the same time, one has \( AE_t < PE_t \) and surplus = 0. Under this condition, \( AE_t \) is calculated using equation (14) below

\[
AE_t = PE_t \cdot \frac{W_{t-1}}{FC}
\]

(14)

In this study two modifications are made. In the original model the potential evapotranspiration, \( PE_t \) is calculated using the temperature based Thornthwaite equation, which is an empirical and site-specific estimation. In this study it is calculated using the Penman-Monteith equation which is a physically based model that proved to be more reliable in various parts of the world (Allen et al. 1998). The second modification is towards the calculation of actual evapotranspiration using equation (14). In arid climates, especially during the dry and transit seasons, soil moisture is generally very low, and as a consequence, the value of \( W_{t-1}/FC \) can be very small (nearly zero). Equation (14) may therefore significantly underestimate actual evapotranspiration especially in a dry season with occasional small rain. To overcome this drawback, actual evapotranspiration was calculated using a modified equation (14a).

\[
AE_t = \min \left\{ PE_t, P_t + PE_t \cdot \frac{W_{t-1}}{FC} \right\}
\]

(14a)

In practice, if the initial soil moisture is unknown, which is typically the case, a balancing routine is used to force the net change in soil moisture from the beginning to the end of a specified balancing period (N time steps) to zero. To do this, the initial soil moisture is set to the water-holding capacity and budget calculations are made up to the time period (N+1). The initial soil moisture \( (W_1) \) at time 1 is then set to be equal to the soil moisture at time N+1 and the budget is re-computed until the difference \( W_1 - W_{N+1} \) reduces to a certain value that is less than a specified tolerance.

One of the parameters that has to be specified is the field capacity (FC) of soil. FC is also known as the soil water holding capacity. It represents the amount of water remaining in the soil after the soil layer has been saturated and the free or drainable water has been allowed to drain away, i.e. after a few days of substantial rain event(s). The soil water holding capacities in the study region are approximately 205, 108, and 154 mm in 1 meter deep soil for stations Abu Naama, Damazine and Gedarif, respectively (available from National oceanographic and atmospheric administration NOAA). In this study this method is used as a reference method, because it is based on the water balance calculation.
Based on energy balance, Bouchet (1963) derived the complementary relationship between actual and potential evapotranspiration; it states that as the surface dries the decrease in actual evapotranspiration is accompanied by an equal, but opposite, change in the potential evapotranspiration. This relationship is described as:

$$AE + PE = 2 \text{ETW} \quad (15)$$

Where $AE$, $PE$ and ETW are actual, potential and wet environment evapotranspiration, respectively. The complementary relationship has formed the basis for the development of several evapotranspiration models (Morton 1983; Brutsaert and Strickler 1979; Granger and Gray 1989), which differ in the calculation of $PE$ and ETW. $AE$ is usually calculated as a residual of (Eq. 1). Complementary relationship calculation methods are now increasingly used to calculate actual evapotranspiration because they require only observations on climate variables and bypass complex and poorly understood soil-plant processes (Xu and Singh 2005). Previous studies (Xu and Chen 2005) showed that the model developed by Granger and Gray (1989) is a good choice.

Considering the general case of evapotranspiration from an unsaturated surface at some rate less than the potential, i.e., $0 < AE < PE$, Grange and Gray (1989) defined relative evapotranspiration, the ratio of actual to potential evapotranspiration, $G = AE / PE$, as a unique parameter for each set of atmospheric and surface conditions. In other words, the surface conditions play a predominant role in the partitioning of the energy available for evapotranspiration, and thus also control potential evapotranspiration. For a wet surface, $G$ will be equal to unity; while for a very dry surface, $G$ will approach zero. Based on the above consideration, Grange and Gray (1989) modified Penman's (1948) energy balance equation for calculating potential evapotranspiration into equation (16) to calculate the actual evapotranspiration:

$$AE = \frac{\Delta G}{\Delta G + \gamma \frac{R_n}{\lambda}} + \frac{\gamma G}{\Delta G + \gamma} E_a$$

where $R_n$ is the net radiation near the surface, $\lambda$ is the latent heat, $\Delta$ is the slope of the saturation vapour pressure curve at the air temperature, $\gamma$ is the psychometric constant, and $E_a$ is the drying power calculated as follows:

$$E_a = f(U_z)(e_s - e_a) \quad (17)$$

Where $f(U_z)$ is some function of the mean wind speed at a reference level $z$ above the ground; and $e_a$ and $e_s$ are the actual vapour pressure of the air and the saturation vapour pressure at the air temperature, respectively. Penman (1948) originally suggested an empirical linear approximation for $f(U_z)$ which is used here:

$$f(U_z) = 0.0026 (1 + 0.54 U_z) \quad (18)$$

Granger and Gray (1989) further derived that the relative evaporation $G$ can be calculated as:

$$G = \frac{1}{1 + 0.028 e^{-0.045 D}}$$

(19)
Where $D = E_a / (E_a + R_a / \lambda)$ is the relative drying power. Granger (1998) modified equation (19) to

$$G = \frac{1}{0.793 + 0.2e^{4.902 D} + 0.006 D}$$

(20)

Previous studies (Xu and Chen, 2005; Xu and Singh 2005) showed that the constants 0.793 and 0.2 in equation (20) need to be locally calibrated to improve the estimation by the GG method, and suggested a general form as

$$G = \frac{1}{a_2 + b_2 e^{4.902 D} + 0.006 D}$$

(21)

Where $a_2$ and $b_2$ are considered as parameters to be calibrated.

In order to calibrate parameters $a_2$ and $b_2$ to give the best estimate of actual evapotranspiration using equation (16), the total annual actual evapotranspiration data is needed, which is not available in the study region. Annual evapotranspiration is thus calculated based on the Budyko-type framework as a function of the aridity index. Different empirical equations of this type exist (e.g., Dooge, 1992; Xu and Singh, 2004) to describe this relationship (i.e., Schreiber 1904; Ol’dekop 1911; and Pike 1964), which are expressed, respectively, as

$$AE = \frac{P}{PE} [1 - \exp(-\frac{PE}{P})]$$

(22)

$$AE = \tanh(\frac{P}{PE})$$

(23)

$$AE = \frac{P}{PE} \sqrt{1 + (\frac{P}{PE})^2}$$

(24)

Where $AE$, $P$, and $PE$ are the annual actual evapotranspiration, precipitation, and potential evapotranspiration as calculated by Penman-Monteith equation (Allen et al. 1998) respectively. In this study the average value calculated by these three methods is used as the annual total actual evapotranspiration for calibrating parameters $a_2$ and $b_2$ in the GG method.
4. Results and Discussion

4.1. Paper I: Understanding the erosion process in terms of natural factors and identification and estimation of eroded areas

ASTER DEM generation and orthorectification

Two DEMs were extracted from the two ASTER images and were compared to SRTM. The results (Figure 4a and 4b, Paper I) showed that the two ASTER-generated DEMs suffered vertical errors and were less accurate than SRTM. One reason can be the fact that the study area is predominantly a rural area with few land cover/use classes, therefore it is characterised by low optical and radiometric image contrast. Also the accuracy of the ground control points (GCPs) was among the main factors limiting the accuracy of ASTER DEM (Table 2, Paper I). Compared to the SRTM, ASTER DEM systematically overestimated elevation at higher latitudes while it underestimated elevation at low altitudes (Figure 5, Paper I). The maximum errors occurred at sharp ridges or deep gullies. Accordingly SRTM was used as the DEM source in this study.

Local and regional identification and estimation of soil erosion

The bi-temporal two scale soil erosion model which was developed in the study, was used to map the spatial distribution of gully erosion in the Blue Nile region during 2006 seasonal rains. A regional spatial distribution map of seasonal gullies in the Blue Nile was produced (Fig. 5-1). It is seen in the figure that gully erosion is evident in and around the Blue Nile River and its tributaries. The bi-temporal classification of ASTER images resulted in an estimation of the increase in the area of gullies. The increase is estimated at approximately 112 km$^2$ (8.7%) of the total image area (Table 7, Paper I) during the rainy season of 2006. The classification of MODIS products allowed the identification and estimation of erosion on a regional scale. During the same period, erosion was estimated at 2071 km$^2$ (6.5%) of the total MODIS area (Table 8, Paper I). The classification of ASTER March was better than ASTER December because VNIR and SWIR channels are more capable of discriminating erosion gullies in the dry season that is characterised by less soil moisture and therefore has higher spectral reflectance. The overall accuracy of MODIS is less than that of ASTER because there is insufficient spectral distinctiveness due to the low spectral and spatial resolution of the MODIS data. Also MODIS products consist of averaged products rather than the actual spectral bands. There were some artefacts in MODIS that interfered with the calculation of eroded areas. The results of this study should be viewed in light of the facts that: (a) the year 2006 was an exceptionally wet year compared to previous years, (b) the Upper Blue Nile River is steeper and receives more rain than any other rivers in Sudan namely, the White Nile, River Atbara, Sobat, etc., and (c) there has not been any similar previous studies reported in this region.
Natural factors affecting gully erosion

The study also found that as natural factors, river flow network and slope were more important in causing erosion in the Blue Nile region in Eastern Sudan than aspect and elevation especially during and after the rainy season (Figures 6a and 6b, Paper I). In December the river networks continue to flow even when the rain has stopped carrying with them eroded soil and weathered rock debris from the hill slopes and valley walls. As slope steepness increase, the amount of water contributed by upslope areas and the velocity of water flow also increase, causing stream power and increasing potential erosion.
4.2. Paper II: Three methods for estimation of regional evapotranspiration

Daily and monthly estimation of actual evapotranspiration

Comparisons of the daily $AE$ (Figure 3, Paper II) and monthly $AE$ (Figures 7a-7c, Paper II) estimates revealed that in the dry season (November – April) and the transit season from dry to wet (May – June), the $AE$ calculated by the SEBAL and GG methods were significantly higher than those by the WB method. This is because during the dry season the soil moisture is very low, hence lowering the WB estimate of $AE$. The other two methods failed to catch this reality and hence they overestimated $AE$. In the wet season (July-October), there is generally better agreement between the SEBAL method and the WB method than that between the GG method and the WB method, although this agreement varied for particular days and stations (Figure 5, Paper II). The GG method showed an underestimation of daily $AE$ in the wet season. Generally speaking there was better correlation between SEBAL and WB than between GG and WB estimates. The estimated daily $AE$ followed the trend of soil moisture variation whereas the seasonal pattern of $AE$ was a result of the combined effect of rainfall and soil moisture.

Mapping actual evapotranspiration using SEBAL

Subsequently, SEBAL was used to map actual evapotranspiration in the region (Figure 5-2: a-l). The figure revealed that during the dry period (j, k, l, a, b, c, and d) the north-east of the Blue Nile exhibits low levels of $AE$ which is approximately 0-20 mm month$^{-1}$, because this region is bare land and very dry at this time of the year. During the same period, the southern region which is between the Blue and the White Nile rivers showed higher rates of $AE$ as indicated by the yellow/green colour correspondent to $AE \approx 30-90$ mm month$^{-1}$. The highest evapotranspiration rates were found in north-western region between the White and the Blue Niles where the Gezira scheme is located which resulted from the irrigation activities. However, during the wet period (e, f, g, h, and i), and without the influence of irrigation activities, a different picture was found. The highest $AE$ values ($AE \approx 90-150$ mm month$^{-1}$) were found in the south-east part of the study region, and were decreasing towards the north-west. This was a reflection of the spatial distribution of rainfall in the study region. The rainfall pattern is such that the south-eastern area has more rain during the rainy season than the north-western area (El Tom, 1975). The spatial distribution of $AE$ was a combined result of the amount of rainfall and the availability of soil moisture, which played different roles in different seasons. In conclusion, the spatial distribution maps of monthly $AE$ produced by the SEBAL method show that in the dry season, the spatial distribution pattern is determined by the location, aspect, land use and irrigation activities. During the wet season, the spatial distribution pattern of monthly $AE$ follows that of the rainfall distribution.
Figure 4-2 (a-l) Spatial distribution map of monthly actual evapotranspiration

(This figure is re-produced from Paper II)
4.3. Paper III: Understanding rainfall, temperature and soil moisture trends and relationships plus identifying soil moisture variability in terms of regeneration demands.

Fluctuation regimes of soil moisture in NCEP re-analysis 1 data in Sudan

In Sudan as was pointed out in Paper II, the soil moisture pattern follows that of the rainfall very closely. When rainfall increases, soil moisture also increases due to an increase in surface evapotranspiration without any significant change in the atmospheric moisture convergence (Douville et al., 2001). The soil-precipitation two-way interaction is commonly referred to as a positive feedback, since the water added to the land surface during a precipitation event leads to increased evapotranspiration, and this in turn can lead to further rainfall. In order to understand the nature of this soil moisture variability, not only in relation to rainfall, but also in relation to temperature changes, the Mann Kendall test was used.

Annual trends

During the period 1965-2005 soil moisture and rainfall decreased while temperature increased significantly using MK test at 95% confidence interval (Figure 3, Paper III). These trends are detected in the annual series as well as in wet season when the seasonal series was analysed (Figure 4, Paper III). These results are consistent with previous studies (e.g., Alvi, 1994; Janowiak, 1988; Nicholson et al., 2000). Changes in soil moisture are more dominated by changes in temperature than in rainfall. Although rainfall amounts decreased all over the country, only significant decline was noted in the south eastern parts as Osman and Shamseldin (2002) had pointed out also in their study. Generally higher month-to-month variability had characterized the temperatures, with significant increases in the northern and central parts. The observed change in temperature could be partly a consequence of local climatic and human forces, such as change in rainfall pattern (periodic drought), rapid increase in population and the mismanagement of land and water resources, desert encroachment and soil erosion, the interaction of which could modify the local climatic conditions (Hulme, 1989; Hulme and Kelly, 1993). The rainfall amounts decreased in all months (Figure 4, Paper III). In August, the rainfall decreased significantly (solid red inverted triangles) in the southern parts. Trilsbach and Hulme (1984) also reported that heavy rains in the rainy season declined more than median and light rains, and the steepest decline occured in August.

During the dry season the soil moisture continued to show significant decreasing trends in all months but the change was less dramatic or softer than the wet season (Figure 4, Paper III). January temperatures decreased significantly. This affected the western and southern parts of Sudan. February temperatures remained unchanged except for an increase in three stations in the South. March temperatures were a mixture of both decreasing (west) and increasing (south and east) significant trends. During December-March soil moisture continued to decrease despite the unchanged or the decreasing temperatures during the same period. This was attributed partly to other unconsidered factors and to a slight decrease in rainfall. Rainfall decreased in all months although not significantly.

In order to quantitatively evaluate the impact of the rainfall and temperature changes on the decline of soil moisture, the Geographically Weighted Regression (GWR) was used (Figure 5, Paper III). The GWR performed differently in different months and different regions in reflecting the relationship between SM, rainfall and temperature. The GWR performance was reflected by the local estimate of the coefficient ($R^2$). In northern Sudan and because of the
shortages in rainfall, there is no evidence of correlation ($R^2$ ranged between 0.0 to 0.3). During dry months, under the influence of dry north-easterlies, there is practically no rainfall countrywide except for a small area in north-eastern Sudan by the Red Sea. However, during the rainy season July, August, September and October, South Sudan also showed no correlation. The reason could be because of the heterogeneity across the south in terms of physical characteristics (soils, topography, and microclimate) that affect the relationships. For example the rainy season, which increases in length from north to south, has a great effect on the temperature regimes. During the rainy season, temperature drops significantly in the south, where the dry season is short, thus making the peak wet months cooler than the winter months (UNEP, 1992; Hare and Ogallo, 1993). Therefore the GWR relationship does not hold in such a case. In central Sudan, the temperature moderates during the wet season and hence central Sudan shows weak to strong correlations ($R^2$ 0.3-0.6 and above). During the rainy season this relationship is stronger. The SM predicted by the GWR both in time and space was compared to NCEP/NCAR SM data (Figure 5, Figure 6, Paper III). For the prediction of SM data in time, the data set of monthly time series is split into two. The data set for the period 1965-2004 was used for running the GWR model, and the data set of the year 2006 is used as a control for the verification of the model. The result indicated that the model predicted the SM well ($R^2 = 0.9$). For predicting SM in space, the model was used to predict SM in Ad’Damazine station located in south eastern Sudan (at longitude 34.4° W and latitude 11.8° N). The model was fed with observed data of mean monthly rainfall and temperature from the year 2006. The relationship between predicted and “observed” data for Damazine was strong ($R^2 = 0.8$). The ability of GWR to predict both in space and time makes it a strong competitor to traditional methods of regression analysis like Ordinary Least Square (OLS).

**Aridity index and irrigation requirements**

Long-term average AI was calculated for the period 1965-2005. AI <20 is shown in yellow and AI> 20 is shown in green colour (Figure 5-3). This long-term AI was then compared to the average AI of the last 10 years, 1995-2005 with AI> 20 (shown as horizontal lines). During the dry season the AI is below 20 for all the country (except a smaller patch in the south) during both periods. There is therefore very little or no-change in the AI for the past 40 years because there is no rainfall during this season. During the wet season May, July and September the last ten years of the study period showed more areas with AI >20 (horizontal lines). This was attributed to increases in rainfall that were experienced since the mid nineties (Elagib and Mansell, 2000, Hulme et al., 2001). The long term AI>20 for the whole period (1965-2005) was lower probably because of the reported decline in rainfall especially in semi-arid central Sudan since 1965 and which continued and intensified in the 1980s, with 1984 the driest year on record (Walsh, et al., 1988 and Hulme, 1990). When compared to NDVI data, the aridity index AI, performed well in the wet season in reflecting the wet conditions in Sudan.
Figure 4-3 Seasonal distribution of Aridity Index. Two periods are compared: 1) 1965-2005 period a long term average AI < 20 is shown in yellow colour, whilst a long term average of AI > 20 are shown in green colour. 2) The period 1995-2005, a ten years average AI > 20 are shown as black horizontal lines which are superimposed on top of the map of the 1965-2005 period. (This figure is re-produced from Paper III)
5. Conclusion and implications

The vegetation-soil moisture relationships describe the various vegetation patterns which occur in the country. These patterns occur on a variety of soils developed from different parent materials. Often the equilibrium between the vegetation and the environment is delicate and easily displaced due to climate change or interference of man, leading to the destruction of the pattern and the establishment of more xerophytic vegetation. Natural regeneration of Sudan’s vegetation remains the only possible solution for combating this degradation and ultimately contributing to the country economic and social stability. In light of these facts the current thesis tries to understand the physical circumstances that impact vegetation regeneration via studying the connections between soil water, erosion and some of the main elements of the hydrological cycle, namely evapotranspiration, temperature and rainfall. The studies of soil moisture and its climatic and degradation associations in Sudan are rare especially in the thorough way that is presented here. These results have important food security implications, informing agricultural development, environmental conservation, and water resource planning.

This research work addressed the problem of soil erosion and identified erosion gullies and subsequently estimated the eroded area. The research results identified river flow network and slope as more important natural factors, in causing erosion in the Blue Nile region than aspect and elevation. An evaluation of evapotranspiration as a direct reflection of the dynamics of soil moisture and hence vegetation regeneration resulted in the production of a sequence of spatial distribution maps of seasonal actual evapotranspiration. From the maps it was concluded that in the dry season, the spatial distribution pattern is determined by the location, aspect, land use and irrigation activities. In the wet season, the spatial distribution pattern followed that of the rainfall distribution. The seasonal pattern of actual evapotranspiration was a result of the combined effect of rainfall and soil moisture. The seasonal patterns of monthly soil moisture followed those of the monthly rainfall, but there was about one month delay in the phase with the peak of the monthly rainfall appearing in July for all three stations while the peak of the monthly soil moisture appearing in August/September.

An evaluation of the soil moisture, temperature and rainfall variability concluded that:

- There are decreasing trends of SM in Sudan on an annual and seasonal level and that this trend is less dramatic or weaker in the dry season (November-April) than the wet season (May-October).

- Soil moisture variability follows closely that of rainfall and temperature. The results also show that SM variability followed temperature changes more closely than rainfall.

- The spatio-temporal variability of the aridity index, AI, showed that the long-term average of AI is affected by the widely reported decline in rainfall during 1965-1985. The decadal AI average of 1995-2005 gave evidence of increases in rainfall that have been reported since the mid-nineties. During the wet season, the aridity index AI, performed well in reflecting the wet conditions in Sudan.
The evaluation procedure and results presented in this thesis provide valuable information and references for better understanding of soil moisture variation in Sudan and its implication for vegetation regeneration. We have made some exploration of how to assess the irrigation requirements under arid, semi-arid and semi-humid environments and the results could serve as an important index for a comprehensive appraisement of the ecosystem and for sustainable development.

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Identification and mapping of soil erosion areas in the Blue Nile, Eastern Sudan using multispectral ASTER and MODIS satellite data and the SRTM elevation model

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Abstract. The area of the Upper Blue Nile in Eastern Sudan is considered prone to soil erosion which is an important indicator of the land degradation process. In this study, an erosion identification and mapping approach is developed based on adaptations to the regional characteristics of the study area and the availability of data. This approach is derived from fusion between remote sensing data and geographical information systems (GIS). The developed model is used to map the spatial distribution of soil erosion caused by the rains of 2006 using automatic classification of multispectral Advanced Spaceborne Thermal Emission and Reflection Radiometer (ASTER) imagery. Shuttle Radar Topography Mission (SRTM) digital elevation model is used to orthoproject ASTER data. A maximum likelihood classifier is trained with four classes, Gully, Flat land, Mountain and Water and applied to images from March and December 2006. Validation is done with field data from December and January 2006/2007. The results allow the identification of erosion gullies and subsequent estimation of eroded area. Consequently, the results are up-scaled using Moderate Resolution Imaging Spectroradiometer (MODIS) products of the same dates. Because the selected study site is representative of the wider Blue Nile region, it is expected that the approach presented could be applied to larger areas.

1 Introduction

This paper is the second (the first is Xu et al., 2009) in a set of studies to evaluate the spatial and temporal variability of soil water in terms of natural factors as well as land-use changes as fundamental factors for vegetation regeneration in arid ecosystems in the Blue Nile, Eastern Sudan. This study is concerned with evaluating the spatial distribution of soil erosion as one of the implications of soil degradation in the study region. Such an evaluation is regarded as an important initial inventory in order to further assess the soil water content status (Gomer and Vogt, 2000). Erosion can be a result of anthropogenic disturbance such as overgrazing, soil crust disturbance, and climatic changes such as precipitation increases. Or it can happen as a natural process. The important natural factors controlling erosion are, among others, rainfall regime, vegetation cover, terrain, slope, aspect and river flow network (Linsley et al., 1982). Fluctuating rainfall amounts and intensities have significant impacts on soil erosion rates. Where rainfall intensity increases, erosion and runoff increase at an even greater rate. On the other hand, decreasing annual rainfall triggers system feedbacks related to the decreased biomass production that lead to greater susceptibility of the soil to erode (Nearing et al., 2004). Because of the important role of direct rain drop impact, vegetation provides significant protection against erosion by absorbing the energy of the falling drops and generally reducing the drop sizes, which reach the ground. Vegetation may also provide mechanical protection to the soil against soil erosion via the
root system. In addition, an adequate vegetation cover generally improves infiltration through the addition of organic matter to the soil. Rates of erosion are greater on steep slopes than on flat slopes. The steeper the slope the more effective is splash slope erosion in moving the soil down slope. Overland flow velocities are also greater on steep slopes, and mass movements are more likely to occur in steep terrain. The length of slope is also important. The shorter the slope length, the sooner the eroded material reaches the stream, but this is offset by the fact that overland-flow discharge and velocity increase with length of slope. River flow networks are multiple-branching systems, beginning with tiny rivulets flowing downhill during rainstorms that join into rills and gullies and eventually into creeks and streams. In their headwater regions, river networks are primarily erosional. They acquire soil and weathered rock debris from hill slopes and valley walls. Any abrupt change in any part of the system will propagate uphill across the landscape as well as downstream through these drainage networks (Ritter et al., 2002). The river flow network is a representation of the flow accumulation, also known as contributing area. This is the size of the region over which water from rainfall can be aggregated. As specific catchment area and slope steepness increase, the amount of water contributed by upslope areas and the velocity of water flow increase, hence stream power and potential erosion increase (Moore et al., 1988). The influence of these natural factors on soil erosion is the main concern of this paper.

Arid and semi-arid regions are particularly susceptible to soil erosion due to their low plant cover (Bull, 1981). In time, such soil erosion also results in the loss of soil nutrients, particularly carbon and nitrogen (West, 1991; Mokwunye, 1996), due to the restrictions of feedbacks in carbon and nitrogen cycles between plants, atmosphere and soil (Schlesinger et al., 1990). Considerable literature (Thornes, 1990; Ayoub, 1998; Nakileza et al., 1999; Symeonakis and Drake, 2004) point to the widespread natural resource degradation especially in sub-Saharan Africa, including Sudan. The area of the Upper Blue Nile in Eastern Sudan is considered prone to degradation by Symeonakis and Drake (2004).

The objectives of this study are (1) to improve our understanding of the erosion process in the Upper Blue Nile in Eastern Sudan in terms of natural factors, (2) to identify the erosion areas and estimate their changes using multispectral satellite images from the Advanced Spaceborne Thermal Emission and Reflection Radiometer ASTER sensor together with field data, and (3) to upscale and map the spatial distribution of soil erosion area changes in a larger region using Moderate Resolution Imaging Spectroradiometer MODIS.

To achieve these objectives, an erosion identification and mapping approach is developed based on adaptations to the regional characteristics of the study area and the availability of data. This approach is derived from fusion between remote sensing data and geographical information systems (GIS).
2 Study area and data

2.1 Description of study area

The study area lies between latitudes 11° and 16° N and longitudes 33° and 35° E (Fig. 1). It portrays different land use types including agriculture, forests and range areas. Agriculture forms are small holdings, mechanized and river bank farming. The area exhibits high land cover variability. It includes for example the Ar Roseris power dam and its lake. There are a number of forests for example the Okalma forest reserve which is bound by Jabal Okalma (Jabal is the Arabic word for mountain) on the west and Jabal Zign (400 m a.s.l.) on the south east corner. Both mountains constitute a source of sheet floods towards the Okalma forest reserve. The forest reserve is a natural forest composed of a mixture of trees, mainly Acacia seyal and Balanites aegyptiaca plus other species. There is a large variation of age groups from young new regeneration to large groups of forest stands. The forest is related to the surrounding societies providing diversified benefits from the trees and the land. Regeneration and forest development factors are evident. Other areas are abandoned fallow land formed following abandonment of agricultural cropping. To the south-east there is a seasonal Khor Dunya (a gully or a seasonal water course is locally known as khor), running from south west to north east until it joins the Blue Nile. This khor is characterised by the presence of degraded areas, natural regeneration, human activities such as agriculture, pastoralism and small settlements, as well as human intervention to reclaim the vegetation. Gully erosion stripes off the fertile clay soils from the degradational clay plain forming bad-lands known locally as “Kerib” (Mirghani, 2007).

Observed data indicate that the mean annual precipitation in 2006 was 900 mm, of which around 90% is collected during the rainy season (April–October). For erosion we are interested in the inter-annual and not the annual total of rainfall. The maximum two or three rainy days in 2006 are of interest because it is the maximum rainfall that causes erosion. The maximum rainfall occurred on 8 and 10 May with precipitation levels as high as 109.2 mm and 200.6 mm, respectively. The mean annual temperature is 28.8 °C. The lowest daily mean temperature is 13 °C and is measured in December. The highest mean daily temperature is 43.8 °C and is measured in May. The runoff coefficient is 20–30% (Ahmed et al., 2006). The year 2006 was an exceptionally wet year. The normal mean annual rainfall is 500 mm (Abdulkarim, 2006).

2.2 Economic impacts of gully erosion in the study region

Although erosion in the centre of the gully is visually apparent (Fig. 2), its effects are not always detectable in terms of changes in soil quality (Ward et al., 2001). This indicates how geomorphologically-apparent desertification (Nir and Klein, 1974; Rozin and Schick, 1996) and changes in soil nutrient content are not necessarily congruent. Nonetheless, a decline in soil nutrients is recorded as some of the most important soil variables, and thus is likely to significantly impact plant growth. Although the major erosion is usually in the centre of the gully in a strip that is only 5–30 m wide, most plant biomass and species diversity are as well concentrated there (Ward and Olsvig-Whittaker, 1993). Moreover, the concentration of the water current in the central erosion gully necessarily reduces the water flow to the adjoining sides of the valley during floods. As a consequence of this reduction in water availability, leaching of salts is reduced (Shalhevet and Bernstein, 1968; Dan et al., 1973; Dan and Yaalon, 1982; Dan and Koyumdjisky, 1987) and soil salinity increases on the sides of the valley. Thus, even in the soil that remains un-eroded, soil quality declines over time.

2.3 Data selection

Nowadays there is a vast reservoir of remote sensing data, some of them are freely available and easily downloadable from the internet. Remote sensing data are described in terms of spatial resolution, temporal resolution, timing, section of the electromagnetic spectrum, stereo, interferometric or ranging capability, and usability (Kääb et al., 2005). Two types of satellite images are used in this study: Advanced Spaceborne Thermal Emission and Reflection Radiometer (ASTER) and Moderate Resolution Imaging Spectroradiometer (MODIS). In addition, a digital elevation model (DEM) provided by Shuttle Radar Topography Mission (SRTM) was used. Table 1 summarises the data characteristics. The software used are PCI Geomatica version 10.1 and ArcGIS version 9.3. In the following sections each data set is described in detail.
Table 1. Description of the data.

<table>
<thead>
<tr>
<th>Name</th>
<th>Spatial resolution</th>
<th>Temporal resolution</th>
<th>Swath</th>
<th>Used bands</th>
<th>Date</th>
</tr>
</thead>
<tbody>
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<td>ASTER-Terra</td>
<td>15 m</td>
<td>Daily</td>
<td>60×60 km</td>
<td>VNIR (1,2,3)</td>
<td>30 Mar 2006</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>SWIR (5,6,7,8,9)</td>
<td></td>
<td>18 Dec 2006</td>
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<tr>
<td>MODIS-Terra</td>
<td>500 m</td>
<td>15-day average</td>
<td>180×180 km</td>
<td>1,2,5,6,7,8</td>
<td>22 Mar 2006</td>
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<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>19 Dec 2006</td>
</tr>
<tr>
<td>SRTM</td>
<td>90 m vertical accuracy</td>
<td></td>
<td></td>
<td>10–20 m</td>
<td>11 Feb 2000</td>
</tr>
</tbody>
</table>

2.3.1 ASTER data

ASTER is a medium-resolution multispectral satellite sensor on the National Aeronautics and Space Administration (NASA) Terra spacecraft, incorporating also an along-track stereo sensor. The latter has stereo angle of about 28° directed backwards. ASTER has 3 NADIR cameras with bands 1–3 in the visible and near infrared (VNIR), bands 4–9 in the short-wave infrared (SWIR), bands 10–14 in the thermal infrared (TIR). It also has a back-looking stereo camera (band 3B). The spatial width of the seasonal gullies in the study area vary between 2 m to 20 m. Therefore a combination of ASTER’s VNIR and SWIR channels of 15 m spatial resolution can capture these phenomena. ASTER thermal infrared bands (TIR) bands (B10-B14) whose spatial resolution is 90 m are therefore not used in this study. Moreover, the reasonable image swath of 60 km allows gully identification over wide regions. The raw data is retrieved in HDF (hierarchical data format). It contains image data, ancillary data (date, time, orbits, positions, angles, sensor and satellite settings, etc.), and meta-data. Terra is preferred to Aqua because it registers images during the day. ASTER level 1B destriped data is used. The two ASTER images are from 30 March 2006 and 18 December 2006. March on the one hand marks the end of the dry season. The rainy season usually begins in early April. December on the other hand is characterised by high soil moisture conditions since it is just towards the very end of the rainy season. It is also characterised by vigorous seasonal and permanent vegetation growth. The scenes were selected for dates with no cloud cover.

2.3.2 MODIS data

At a continental or global scale, coarse spatial resolution data such as from MODIS are preferable because they cover much larger spatial scales simultaneously. MODIS Terra 13Q data version v005, level 2B (available at http://glovis.usgs.gov/), is used. Two MODIS images dating to 22 March 2006 and 19 December 2006 are used. The first image is a 16-days average of daily images between the period 14 to 30 March 2006 and the second is a 16-days average of daily images between the periods 2 to 18 December 2006. MODIS products of versions 4 or higher have been validated and approved for scientific research. Each image has 11 data layers originally. Of these 11 layers, only 6 data layers are used. The layers used are: 1, 2, 5, 6, 7 and 8 which correspond to Normalised Difference Vegetation Index (NDVI), Enhanced Vegetation Index (EVI), red reflectance, Near Infra Red (NIR) reflectance, blue reflectance and Medium Infra Red (MIR) reflectance respectively. The NDVI and EVI are selected because they demonstrate a good dynamic range and sensitivity for monitoring and assessing spatial and temporal variations in vegetation amount and condition. Whereas the NDVI which is chlorophyll sensitive is useful as a vegetation measure in that it is sufficiently stable to permit meaningful comparisons of seasonal and inter-annual changes in vegetation growth and activity. The strength of the NDVI is in its rationing concept, which reduces many forms of multiplicative noise (illumination differences, cloud shadows, atmospheric attenuation, and certain topographic variations) presents in multiple bands. EVI is more responsive to canopy structural variations, including Leaf Area Index (LAI), canopy type, plant physiognomy, and canopy architecture (Gao et al., 2000). The two vegetation indices (VIs) complement each other in global vegetation studies and improve upon the detection of vegetation changes.

ASTER and MODIS are complementary in resolution, offering a unique opportunity for scale-related studies (Vrieling et al., 2008). ASTER with its finer spatial resolution and better accuracy is used to identify erosion areas and to quantify the erosion for the small area, while MODIS is used to upscale ASTER results for sake of understanding erosion on a larger scale for the wider Blue Nile region in Eastern Sudan.

2.3.3 SRTM data

Shuttle Radar Topography Mission (SRTM) is a single-pass InSAR, which provides elevation data on a near-global scale (between 60° N and 54° S), and it is the most complete high-resolution digital topographic database of Earth. It has gaps in it that resulted from shadow, SRTM3 is freely available. Where available, the SRTM indeed represents a revolutionary data set for all kinds of terrain studies (Kääb et al., 2005).
3 Methodology

3.1 Method overview

The flow diagram in Fig. 3 demonstrates the method used in developing this erosion area model. Briefly speaking, the study area (Fig. 1) is divided into two scales:

1. A 60 × 60 km (approx. 3600 km²) scale. This area is studied in detail with the aid of the finer resolution of two ASTER images dated 30 March and 18 December of the year 2006. The aim is to detect the bi-temporal changes in eroded area that have resulted from the seasonal rains of 2006. An initial preprocessing of the raw data is undertaken. ASTER images are then used to generate photogrammetric DEMs. The qualities of these DEMs are then compared to SRTM. The best DEM of the three is hence used to accurately orthorectify ASTER images. The orthoimages are then trained using a supervised classifier into four different land cover classes; Gully, Flat, Mountain and Water. The outcome of the classification is validated against field data. Finally, the change in eroded area between the two scenes is estimated. As such the erosion area between the period March and December 2006 is determined at this scale.

2. For up-scaling the verified ASTER results to an 180 × 180 km (approx. 32 400 km²) scale, two MODIS images dating 22 March and 19 December 2006 are used. The MODIS images are pre-processed and georeferenced before undertaking a supervised classification using the same land cover classes and training areas as ASTER. The outcome of MODIS classification is validated against field data and finally the seasonal change in erosion area for the whole Blue Nile region is detected.

3.2 Pre-processing: ASTER DEM generation and orthorectification

Due to radar shadow, foreshortening, layover and insufficient interferometric coherence, the SRTM has significant voids in high mountains. Therefore in this study two DEMs are extracted from each one of the two ASTER images. The qualities of these two DEMs are then compared to SRTM and the most accurate of the three is used further for orthoprojection of ASTER images. Orthoprojection is a mandatory pre-processing step necessary to prevent strong topographically induced distortions between the images in the rugged study area which is characterised by elevation ranging approximately from 10 m to 1500 m. DEM generation from satellite imagery uses photogrammetric principles. Toutin (2001, 2002), and Toutin and Cheng (2001, 2002) outlined the main digital processing steps for DEM generation from ASTER within the PCI Geomatica software.

Briefly, the first three (NADIR) channels 123 N and the back looking channel 3B of ASTER 30 March 2006 are read as two separate .pix files in PCI Geomatica’s orthoenhine using Toutin’s low resolution model. The scenes are reprojected to UTM WGS1984 Zone 36 N projection. The ground control points (GCPs) necessary for estimating the model parameters are collected from a hillshade image of SRTM generated in ArcGIS 9.3. The reason for using the hillshade of SRTM for collecting GCPs is due to the lack of accurate topographic maps for the study area. The hillshade SRTM ensures better co-registration between the SRTM and ASTER-derived products later in the combination process between the two. In total 29 GCPs are collected for each image (Table 2). The Root Mean Square Error (RMSE) for NADIR SRTM is 0.98 pixels (≈13.3 m) and for the back-looking 3B is 1.69 pixels (≈23.8 m). The two images 123 N and 3B are tied together using 95 tie points. The maximum residual for tie points is 0.77 pixels (≈8.67 m). Using cubic transformation and SRTM as the DEM source, a 60 m resolution DEM (March DEM) is extracted from 30 March ASTER image.
The same procedure is repeated using the second ASTER image of 18 December. The RMS is shown in Table 2. A second 60 m resolution DEM (December DEM) is also extracted.

### 3.3 ASTER classification

After orthorectifying ASTER images using SRTM, the channels that best describe the process of soil erosion are selected. Considering the spatial width of gullies, ASTER’s VNIR and SWIR channels at 15 m spatial resolution have been analysed. Additional layers that represent some of the most important natural factors causing erosion are added. These layers are elevation, slope, aspect, and river flow network. Therefore the final ASTER orthoimages used for classification consisted of 13 layers stacked together. These include: VNIR channels 1, 2 and 3, SWIR channels 4, 5, 6, 7, 8 and 9, SRTM, slope, aspect, and river flow network. The last three layers; slope, aspect and river flow network are all calculated from SRTM in ArcGIS 9.3. Both slope and aspect are calculated from SRTM using a finite difference method that uses eight neighbours in ArcGIS. The two are then resampled to 15 m before adding them to the layer stack. The river flow network was estimated using the hydrologic analysis package in ArcGIS 9.3 as follows. When working with raster DEMs and computing slopes between grid cells, the ratio of the vertical and horizontal resolutions determines the minimum non-zero slope that can be resolved. In this study, the vertical resolution of the DEM is 1 m and a grid size is 15 m. Therefore the resolvable slope of 1/15=0.07. This value means that slopes on hillsides can be computed with a relatively small error. However, slopes in channels are often much smaller than this value. As a consequence, these areas will appear horizontal in the DEM. Therefore in order to avoid this problem, ArcGIS 9.3 uses an eight direction (D8) flow model. The direction of flow is determined by finding the direction of steepest descent, or maximum drop, from each cell. This is calculated as

$$\text{maximum drop} = \frac{\text{change in z-value}}{\text{distance}}$$

The distance is calculated between cell centres. There are eight valid output directions relating to the eight adjacent cells into which flow could travel. One challenge arises if all neighbours are higher than the processing cell. In such a case the processing cell is called a sink and has an undefined flow direction because any water that flows into a sink cannot flow out. To obtain an accurate representation of flow direction across a surface, the sinks should be filled. The minimum elevation value surrounding the sink will identify the height necessary to fill the sink so the water can pass through the cell. A digital elevation model free of sinks is called depressionless DEM. Using the depressionless DEM as an input to the flow direction process, the direction in which water would flow out of each cell is determined. After determining the flow direction, then the flow accumulation is determined. Afterwards, a stream network is created by applying a threshold value to select cells with high accumulated flow in order delineate the stream network. More details on the technique of deriving flow direction from a DEM and on how to create river network in ArcGIS can be found in Jenson and Domingue (1988) and in Tang and Liu (2008), among others.

After stacking all appropriate layers, then supervised maximum likelihood classification is performed aided by the authors’ knowledge of the area. The image is trained into four land cover types: Gully, Flat_land, Mountain and Water. Water bodies are clear and easy to train considering the fine ASTER resolution for both periods of the year. Mountains are easily classified with the aid of the DEM. However, the most challenging task is to separate between Gully and Flat_land since these two classes are bound to overlap and overlapping training area boundaries reduces the reliability of the training sites. To avoid this kind of overlap, the following steps were taken

1. MODIS vegetation indices (VI) are used as auxiliary data to discriminate between training classes; Gully and Flat_land. At first MODIS NDVI and EVI signatures are used to discriminate between two classes; stable vegetation and unstable vegetation. The unstable vegetation is seasonal vegetation that grows during the rainy season; this vegetation is flushed away with erosion, indicating that areas where there is unstable vegetation there is also erosion. The stable vegetation on the other hand is there throughout the year hence no erosion. Where NDVI and EVI values are low, this is an indication of limited, unstable vegetation hence higher erosion risk.

<table>
<thead>
<tr>
<th>Image</th>
<th>Channel</th>
<th>GCP</th>
<th>RMSE X</th>
<th>RMSE Y</th>
</tr>
</thead>
<tbody>
<tr>
<td>ASTER March</td>
<td>Nadir 123 N</td>
<td>29</td>
<td>0.98 pixels (13.3 m)</td>
<td>0.81 pixel (12.00 m)</td>
</tr>
<tr>
<td>ASTER March</td>
<td>Back-looking 3B</td>
<td>29</td>
<td>1.69 pixels (23.8 m)</td>
<td>0.87 pixel (11.90 m)</td>
</tr>
<tr>
<td>ASTER December</td>
<td>Nadir 123 N</td>
<td>31</td>
<td>1.08 pixels (14.6 m)</td>
<td>0.84 pixel (12.06 m)</td>
</tr>
<tr>
<td>ASTER December</td>
<td>Back-looking 3B</td>
<td>31</td>
<td>1.72 pixels (24.2 m)</td>
<td>0.91 pixel (13.0 m)</td>
</tr>
</tbody>
</table>
Higher values of NDVI and EVI indicate more stable vegetation.

2. The layer of river network is superimposed on top of the image to help guide and discriminate between Gullies and Flat_land.

3. Care is taken when selecting the different bands both for display, in either grayscale or as false colour composites FCC.

4. Between each run and the next the classes are refined by varying the number of training areas until better accuracies are achieved.

As many training areas as possible are trained because generally speaking, the more areas identified as training sites, the higher the accuracy of the classification. Once the training areas are defined, then the signature separability values are studied. Signature separability is the statistical difference between pairs of spectral signatures. It is expressed in terms of Bhattacharrya Distance and Transformed Divergence. These are measures of the separability of a pair of probability distributions. Both Bhattacharrya Distance and Transformed Divergence are shown as real values between zero and two. Zero indicates complete overlap between the signatures of two classes; two indicates a complete separation between the two classes. The higher the separability (i.e. more than 1.5) value the more accurate is the classification accuracy. The training areas are tuned until higher signature separability value of 1.5 or more are achieved. After achieving the best accuracies, the signature statistics report is studied in order to determine which channels of the 13 stacked layers are more significant in delineating erosion gullies.

3.4 Post classification

ASTER classification is validated via running an automatic random accuracy assessment. The accuracy assessment is designed and implemented by generating a random sample of 300 points and comparing it to the original orthorectified ASTER image. ASTER image is used as a reference image due to lack of accurate and updated maps for the study area. Each of the 300 samples is assigned to the different classes.

Once the two ASTER images are successfully classified and validated, they are used to study the bi-temporal change. For each image the area represented by each of the four classes is calculated using the function Generate Area Report in PCI Geomatica. The areas in December image are subsequently subtracted from their correspondent in March image in order to calculate the changed area per class.

3.5 Upscaling using MODIS

Once ASTER classification results are accepted, the results are then up-scaled using MODIS. Upscaling allows generalisation of the results of ASTER classification to larger area covering the whole of the Blue Nile region. MODIS is first re-projected to UTM WGS1984 Zone 36 N. Thereafter, it is trained into four classes; Gully, Flat_land, Mountain and Water. MODIS is trained using:

1. MODIS NDVI and EVI are used as auxiliary data to discriminate between training classes; Gully and Flat_land discriminating first between high values i.e. stable vegetation i.e. no erosion and low values i.e. unstable vegetation i.e. erosion areas.

2. The same training areas from ASTER which are converted into shape file and laid over the MODIS image to guide the training.

3. River flow network layer overlaid over the images to guide the training of gullies

MODIS classification is validated against field data in terms of registered digital photos with co-ordinates and time taken in January 2007 from different locations.

4 Results and discussion

4.1 Comparison of the three DEMs

Because the two generated ASTER DEMs cannot be tested against an existing reference DEM, they are tested through the overlay of different orthoimages taken from different sensor positions (also know as multi-incidence angle image). For 30 March ASTER, two orthoimages are produced one for 123 N and another for 3B using March DEM. The two orthoimages are then overlaid to test if they overlap perfectly well pixel-by-pixel using animation flickering techniques. The results showed that the two did not overlap perfectly. The explanation for this is that the used March DEM has vertical errors which translated into horizontal shifts between the orthoprojected pixels. These shifts cause the two multi-incidence images not to overlap perfectly. Next December DEM is tested in a similar manner. The results of the flickering technique showed some vertical errors. It is then concluded that the two ASTER DEMs both suffer vertical errors. Accordingly SRTM is selected as “reference DEM” for all further steps. In order to quantify the error of ASTER March and December DEMs, their contour lines are compared to those of SRTM (Fig. 4a and b, respectively). The depicted test area in the figure represents rugged high-mountain conditions with elevations of up to 1500 m a.s.l., steep rock walls, deep shadows that are without contrast. Therefore, the test area is considered to represent a worst case for DEM generation from ASTER data.

Figure 4a and b shows that the contour lines of two ASTER DEMs (dotted lines) are different to those of the SRTM (solid lines) and that this difference increases with elevation. Figure 4b shows that December DEM has more
deviation from SRTM than March DEM (Fig. 4a). The relative error of the two ASTER DEMs from the “reference” SRTM is plotted and shown in Fig. 5. Figure 5 shows that up to 400 m elevation, the two DEMs have lower values than SRTM. They are also similar in magnitude. After 400 m the two DEMs then have higher values than SRTM at higher altitudes.

The overall conclusions of the DEM comparisons are:

– Using orthoimage overlay techniques for multi-incidence angle images, it is found clearly that the two ASTER-generated DEMs suffer vertical errors and are less accurate because of the low optical and radiometric image contrast resulting from the fact that the study area is predominantly rural with few land cover/use classes. Also the accuracy of GCPs is among the main factors limiting the accuracy of ASTER DEM.

– Compared to the SRTM, ASTER DEM systematically overestimates elevation at higher latitudes while it underestimates elevation at low altitudes. These maximum errors occur at sharp ridges or deep gullies.

– Compared to SRTM December DEM is more accurate up to 400 m elevation, after that it has errors of larger magnitude than March DEM

Accordingly SRTM was used to orthoproject the two ASTER images for classification purposes.

4.2 Factors affecting gully erosion

In order to understand which input channels mostly contributed to gully identification the standard deviation of the signatures for each of the 13 channels for ASTER March and December are plotted in Fig. 6a and b, respectively. Figure 6a and b gives an overview of the contribution of the different channels in the identification of the class Gully. Channels that have high standard deviation values are the least significant in delineating gullies and vice versa. In March which is the dry season (Fig. 6a), river network and slope have the highest standard deviation. In December, however (Fig. 6b),
they have the lowest standard deviation values and hence they are the most important in Gully classification. During and after the rainy season, river network and slope are the most important factors in the erosion process. In December river networks continue to flow even when the rain has stopped carrying with the flow soil and weathered rock debris from hill slopes and valley walls and as slope steepness increase, the amount of water contributed by upslope areas and the velocity of water flow increase, hence stream power and potential erosion increase. As natural factors, river flow network and slope are more important in causing erosion in the Blue Nile region than aspect and elevation.

4.3 Validation of the classification outcome

ASTER classification reports (Table 3) shows an overall accuracy of 82.2% and 75.2% for March and December images, respectively. And similarly for ASTER validation, the accuracy assessment report (Table 4) shows that in March the overall accuracy is 93% and in December is 89%. The classification of March image is better than December because VNIR and SWIR channels are more capable of discriminating erosion gullies in the dry season due to higher spectral reflectance.

MODIS classification accuracies for March and December are 77.2% and 81.0% respectively (Table 5). Automatic accuracy assessment (Table 6) shows overall accuracies of 89% and 88.7% for March and December, respectively. Additionally, the classified MODIS image is validated against a set of thirteen digital photos taken in the field. Nine validation points showed gullies coinciding with the results of the classification (Fig. 7). In March classification is better than that of December because in the dry season there is less soil moisture therefore higher surface reflectance. The overall accuracy of MODIS is less than that of ASTER. The reasons are that there is insufficient spectral distinctiveness due to the low spectral and spatial resolution of the MODIS data; MODIS products consisted of averaged products rather than the actual spectral bands; some artefacts interfered with the calculation of eroded areas.

4.4 Estimation of soil erosion on ASTER scale

The results of the bi-temporal change of gully erosion in ASTER scale are shown in Table 7. Table 7 shows an increase by approximately 112 km² in the area of gullies. This is because after the rainy season, both rain and flood water...
Table 4. Validation of ASTER classification: accuracy statistics report indicating the outcome of the accuracy assessment of the used supervised maximum likelihood classifier.

<table>
<thead>
<tr>
<th>Class name</th>
<th>Producer’s accuracy</th>
<th>User’s accuracy</th>
<th>Kappa statistic</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mountain</td>
<td>88.2%</td>
<td>97.8%</td>
<td>0.97</td>
</tr>
<tr>
<td>Gully</td>
<td>92.2%</td>
<td>92.2%</td>
<td>0.90</td>
</tr>
<tr>
<td>Flat</td>
<td>96.3%</td>
<td>92.3%</td>
<td>0.83</td>
</tr>
<tr>
<td>Water</td>
<td>70.0%</td>
<td>87.5%</td>
<td>0.87</td>
</tr>
</tbody>
</table>

ASTER March

Overall accuracy 93.0%

<table>
<thead>
<tr>
<th>Class Name</th>
<th>Producer’s accuracy</th>
<th>User’s accuracy</th>
<th>Kappa statistic</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mountain</td>
<td>69.2%</td>
<td>75.0%</td>
<td>0.74</td>
</tr>
<tr>
<td>Gully</td>
<td>87.0%</td>
<td>85.3%</td>
<td>0.78</td>
</tr>
<tr>
<td>Flat</td>
<td>95.2%</td>
<td>79.7%</td>
<td>0.75</td>
</tr>
<tr>
<td>Water</td>
<td>91.7%</td>
<td>100%</td>
<td>1.00</td>
</tr>
</tbody>
</table>

ASTER December

Overall accuracy 89.0%

Table 5. Classification report of MODIS products.

<table>
<thead>
<tr>
<th>Image</th>
<th>Average accuracy</th>
<th>Overall accuracy</th>
<th>Kappa coefficient</th>
</tr>
</thead>
<tbody>
<tr>
<td>MODIS March</td>
<td>77.2%</td>
<td>61.9%</td>
<td>0.40</td>
</tr>
<tr>
<td>MODIS December</td>
<td>81.0%</td>
<td>63.8%</td>
<td>0.20</td>
</tr>
</tbody>
</table>

dissects large areas creating either new gullies or increasing the width and/or depth of existing gullies. Naturally an increase in dissected land reduces the extent of flat land. That is why there is a decrease of flat land by about 153 km². The area covered with water has increased by 31.5 km². That is due to the fact that in December there are still large areas that are covered with rain and flood water from the rainy season time which have not receded, percolated or evaporated yet. The total area classified as mountains did not change. In fact this is regarded as an indication of the success of the classification because when stable features like mountains remain stable then this is regarded as an indication of good georeferencing and subsequently good multitemporal analysis (Kääb, 2005).

4.5 Estimation of erosion on a regional scale

The change in the total area of gullies is estimated by subtracting the area in MODIS December image from that in MODIS March (Table 8). Table 8 shows that gullies have increased by approximately 2071 km², and water by 1791 km², respectively. Flat land has decreased by 3864 km² while mountains remained unchanged. Hence the estimated area along the Blue Nile River, Eastern Sudan that was affected by soil erosion during the 2006 rainy season was estimated to be 2071 km² or 6.5% of the total study area approximately. This estimation should be viewed in light of the facts that: (a) the year 2006 was exceptionally wet year compared to previous years, (b) the Upper Blue Nile River is steeper and receives more rain than any other river in Sudan namely, the White Nile, River Atbara, Sobat, etc., and (c) there has not been any previous studies reported in this region.

After close comparison and evaluation of the soil erosion model as presented above, it is used to map the spatial distribution of gully erosion in the Blue Nile region during 2006 seasonal rains (Fig. 7). It is seen in the figure that gully erosion is evident in and around the Blue Nile River and its tributaries.

5 Summary and conclusion

Soil erosion poses a serious environmental and socio-economic threat to the environment and to mankind. Previous research in sub-Sahara Africa has singled out the Upper Blue Nile as an erosion prone area that is recommended for further monitoring and evaluation. In this study a soil erosion area model is suggested. The model benefits from advances in GIS and remote sensing fusion techniques. The model is simple, robust and straightforward. It makes use of the well tested methods of supervised classification using maximum likelihood (MLC). Generalisation of ASTER classification results using MODIS was useful for capturing a regional impression of the spatial distribution of erosion. Key to the model’s success is further development of proper validation procedures. The size of this region makes the traditional mapping methods of aerial photography and field surveying of limited use. Moderate resolution multispectral data allows continental- to global-scale mapping of the Earth’s surface.

Hydrol. Earth Syst. Sci., 14, 1167–1178, 2010

www.hydrol-earth-syst-sci.net/14/1167/2010/
Table 6. Validation of MODIS classification: accuracy statistics report indicating the outcome of the accuracy assessment of the used supervised classifier.

<table>
<thead>
<tr>
<th>Class name</th>
<th>Producer’s accuracy</th>
<th>User’s accuracy</th>
<th>Kappa statistic</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mountain</td>
<td>94.1%</td>
<td>80.0%</td>
<td>0.76</td>
</tr>
<tr>
<td>Gully</td>
<td>90.6%</td>
<td>93.6%</td>
<td>0.91</td>
</tr>
<tr>
<td>Flat</td>
<td>89.7%</td>
<td>89.7%</td>
<td>0.83</td>
</tr>
<tr>
<td>Water</td>
<td>75.0%</td>
<td>90.0%</td>
<td>0.89</td>
</tr>
</tbody>
</table>

Overall accuracy: 89.0%

<table>
<thead>
<tr>
<th>Class name</th>
<th>Producer’s accuracy</th>
<th>User’s accuracy</th>
<th>Kappa statistic</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mountain</td>
<td>91.2%</td>
<td>80.0%</td>
<td>0.74</td>
</tr>
<tr>
<td>Gully</td>
<td>87.6%</td>
<td>92.5%</td>
<td>0.88</td>
</tr>
<tr>
<td>Flat</td>
<td>85.7%</td>
<td>89.9%</td>
<td>0.70</td>
</tr>
<tr>
<td>Water</td>
<td>72.6%</td>
<td>91.4%</td>
<td>0.89</td>
</tr>
</tbody>
</table>

Overall accuracy: 88.7%

Table 7. Local change of erosion area as indicated by the difference (diff) in area percentage (%) or in km² between March and December 2006 for the four land cover classes.

<table>
<thead>
<tr>
<th>Class</th>
<th>Mar Area (%)</th>
<th>Dec Area (%)</th>
<th>Diff</th>
<th>Mar Area (km²)</th>
<th>Dec Area (km²)</th>
<th>Diff</th>
</tr>
</thead>
<tbody>
<tr>
<td>Gully</td>
<td>42.1</td>
<td>50.9</td>
<td>8.77</td>
<td>538.5</td>
<td>650.7</td>
<td>112.2</td>
</tr>
<tr>
<td>Flat land</td>
<td>46.6</td>
<td>34.63</td>
<td>−12.2</td>
<td>596</td>
<td>442.9</td>
<td>−153.1</td>
</tr>
<tr>
<td>Mountain</td>
<td>4.43</td>
<td>4.43</td>
<td>0.00</td>
<td>56.7</td>
<td>56.701</td>
<td>0.05</td>
</tr>
<tr>
<td>Water</td>
<td>6.87</td>
<td>10.07</td>
<td>3.20</td>
<td>67.9</td>
<td>99.49</td>
<td>31.59</td>
</tr>
</tbody>
</table>

Table 8. Regional change of erosion area as indicated by the difference (diff) in area percentage (%) or in km² between March and December 2006 for the four land cover classes.

<table>
<thead>
<tr>
<th>Class</th>
<th>Mar Area (%)</th>
<th>Dec Area (%)</th>
<th>Diff</th>
<th>Mar Area (km²)</th>
<th>Dec Area (km²)</th>
<th>Diff</th>
</tr>
</thead>
<tbody>
<tr>
<td>Gully</td>
<td>33.1</td>
<td>39.7</td>
<td>6.55</td>
<td>10 467</td>
<td>12 538</td>
<td>2071</td>
</tr>
<tr>
<td>Flat land</td>
<td>39.8</td>
<td>27.6</td>
<td>−12.2</td>
<td>12 572</td>
<td>8708</td>
<td>−3864</td>
</tr>
<tr>
<td>Mountain</td>
<td>21.3</td>
<td>21.3</td>
<td>0</td>
<td>6737</td>
<td>6737</td>
<td>0</td>
</tr>
<tr>
<td>Water</td>
<td>5.84</td>
<td>11.5</td>
<td>5.67</td>
<td>1845</td>
<td>3636</td>
<td>1791</td>
</tr>
</tbody>
</table>

while retaining sufficient resolution for geomorphic and ecological studies.

The following conclusions are drawn from the study:

1. ASTER-derived DEMs are less accurate than SRTM because of the low optical contrast of the study area which is predominantly rural with few land cover/use classes low.

2. The incision of gullies found in this area is strongly associated with topography, especially river flow networks and slope.

3. Moreover, on-going climate change can initiate new processes via increased rainfalls, or weaken land cover.

The model can be used to study longer-term changes in erosion by using time series of images from different years. The methodology presented here is also transportable to other arid and semi-arid parts of Sudan where an understanding of soil erosion is desirable.

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References


Spatial and temporal variation of precipitation in Sudan and their possible causes during 1948–2005

Zengxin Zhang · Chong-Yu Xu · Majduline El-Haj El-Tahir · Jianrong Cao · V. P. Singh

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Abstract Temporal and spatial patterns of precipitation are essential to the understanding of soil moisture status which is vital for vegetation regeneration in the arid ecosystems. The purposes of this study are (1) to understand the temporal and spatial variations of precipitation in Sudan during 1948–2005 by using high quality global precipitation data known as Precipitation REConstruction (PREC), which has been constructed at the National Oceanic and Atmospheric Administration (NOAA) Climate Prediction Center, and (2) to discuss the relationship between precipitation variability and moisture flux based on the NCEP/NCAR reanalysis data in order to ascertain the potential causes of the spatial and temporal variations of precipitation in the region. Results showed that (1) annual and monthly precipitation in Sudan had great spatial variability, and mean annual precipitation varied from almost nil in the North to about 1500 mm in the extreme Southwest; (2) precipitation of the main rain season, i.e., July, August and September, and annual total precipitation in the central part of Sudan decreased significantly during 1948–2005; (3) abrupt change points were found in the annual, July, August and September in the late 1960s, when precipitation decreased more rapidly than in other periods; and (4) the decreasing precipitation was associated with the weakening African summer monsoon. The summer moisture flux over Sudan tended to be decreasing after the late 1960s which decreased the northward propagation of moisture flux in North Africa. This study provides a complementary view to the previous studies that attempted to explain the Sahel persistent drought and possible causes.

Keywords Precipitation · Large scale · Abrupt · Moisture flux · Sudan · African monsoon

1 Introduction

In recent years, droughts and subsequent famine and disease have devastated Africa and affected millions of people. With much of rural Africa already struggling to obtain adequate fresh water supplies, the drier conditions and altered precipitation patterns will make meeting the water needs of some of the poorest Africans much harder. Rainfall is the most important water resource in Sudan. However, low levels of rainfall have affected Sudan and resulted in severe droughts in recent years. The years 1983 and 1984 were the most difficult years for the region in recent history (Eltahir 1992). In 1984, crop failure and the spread of water-borne diseases caused by the drought took the lives of 55,000 people which weakened the socio-economic capabilities of the nomadic tribes in Sudan (Osman and Shamseldin 2002).
In Sudan, the rain-fed agriculture produces food for 90% of the population. Rainfall is caused by the southwest monsoon winds flowing from the Atlantic Ocean and the southeast monsoon winds flowing from the Indian Ocean (Osman and Shamseldin 2002). To ascertain the precipitation change and its possible cause in Sudan during the past decades, several observational studies, using meteorological data, point to the climate change in Sudan and these studies confirm that the temperature is rising and rainfall is declining for the past several decades which might accelerate environmental degradation and desertification in Sudan (e.g., Alvi 1994; Janowiak 1988; Nicholson et al. 2000). The precipitation anomaly of Sudan has been related to Sea Surface Temperature Anomalies (SSTAs) in the Gulf of Guinea (Lamb 1978a, b). Palmer (1986) has pointed out that the tropical Indian Ocean SST has a strong influence on the Sahel rainfall. Camberlin (1995) found that the dry and wet conditions over the Sahel were usually associated with warm conditions in the tropical Indian Ocean. Osman and Shamseldin (2002) also investigated the influence of El Niño-southern oscillation (ENSO) and the Indian Ocean Sea surface temperature (SST) on the rainfall variability in the central and southern regions of Sudan and found that the driest years were associated with warm ENSO and Indian Ocean SST conditions. Comprehensive analysis and reviews of rainfall trends and variability in Africa, including the Sahel region and Sudan has been carried out by Hulme and his co-authors (e.g., Trilsbach and Hulme 1984; Hulme 1987, 1990; Hulme and Tosdevin 1989; Walsh et al. 1988). Trilsbach and Hulme (1984) examined rainfall changes and their physical and human implications in the critical desertification zone between latitudes 12°N and 16°N of the Democratic Republic of Sudan and concluded, among others, that heavy falls of rain (>40 mm) are more likely to decline than medium (>10 mm) and light (<10 mm) falls. Walsh et al. (1988) reported that “rainfall decline in semi-arid Sudan since 1965 has continued and intensified in the 1980 s, with 1984 the driest year on record and all annual rainfalls from 1980 to 1987 well below the long-term mean”. Hulme (1990) reported that rainfall depletion has been most severe in semi-arid central Sudan between 1921–1950 and 1956–1985. The length of the wet season has contracted, and rainfall zones have migrated southwards. A reduction in the frequency of rain events rather than a reduced rainfall yield per rain event was found to be the main reason for this depletion.

Increasingly the tendency has been for studies examining the causes of droughts, to look at global scales through the notion of teleconnections. It would seem reasonable that the current drought is a manifestation of an interaction of two or more mechanisms (Cook and Vizy 2006). Nicholson (1986) pointed out that the variations in the Sahel rainfall are generally related to the changes in the intensity of the rainy season rather than to its onset or length as the ITCZ hypothesis would require.

Drought implies some form of moisture deficit and moisture plays an important role in understanding climate change (Prueger et al. 2004). Vertically integrated moisture flux and its convergence/divergence are closely related to precipitation. Much research work has been performed regarding the moisture flux over Africa (e.g., Cadet and Nnoli 1987; Fontaine et al. 2003; Osman and Hastenrath 1969). Cadet and Nnoli (1987) pointed out that the progressive penetration of moisture over West Africa during summer at 850 hPa and the southerly flow penetrates up to 20°N during the maximum activity of the African monsoon. Hulme and Tosdevin (1989) assessed the relationship between tropical easterly jet (TEJ) and Sudan rainfall and found changes in the dynamics and flow of the TEJ exert some control over the Sahelian rainfall. Fontaine et al. (2003) analyzed the atmospheric water and moisture fluxes in West African Monsoon based on NCEP/NCAR and observed rainfall data and found that at more local scales moisture advections and convergences are also significantly associated with the observed Sudan-Sahel rainfall and in wet (dry) situations, with a clear dominance of westerly (easterly) anomalies in the moisture flux south of 15°N.

Although many studies have attempted to explain the Sahel persistent drought in terms of general circulation features, sea surface temperatures, land surface feedback mechanisms, and teleconnected, spatiotemporal rainfall patterns. However, the cause of the Sahel drought remains elusive (Nicholson 1989; Rowell et al. 1995; Long et al. 2000; Trilsbach and Hulme 1984). Long et al. (2000) aimed to advance the understanding of these causes by examining rainfall, horizontal moisture transport, and vertical motion and how they differ regionally and seasonally. There are a few previous studies that have looked at how the precipitation variability is related to the moisture flux in Sudan. The main aim of this study is to provide a complementary view to the previous studies by examining the relationship of moisture flux and precipitation trend in the region. The specific objectives of the study are: (1) to understand the spatial and temporal variability of precipitation from Precipitation REConstructed (PREC) data in Sudan during 1948–2005; and (2) to investigate the relationship between precipitation and moisture flux in the region.

2 Study area and data

Sudan is a vast country with an area of about 2.5 million km² and has an estimated population of about 41.1 million people. The climate ranges from arid in the north to tropical wet-and-dry in the far southwest. About two-thirds
of Sudan lies in the dry and semi-dry region. Temperatures do not vary greatly with the season at any location; the most significant climatic variables are rainfall and the length of the rainy season. Except in the northeast region, from January to March, the country is under the influence of dry northeasterly winds, and during this period there is almost no rainfall countrywide except for a small area in north eastern Sudan where the winds pass over the Mediterranean bringing occasional light rains. By early April, rainy season starts from southern Sudan as the moist south westerlies reach the region, and by August the southwesterly flows extend to the northern limits. The dry north easterlies strengthen to start in September and to push south and cover the entire country by the end of December.

In this paper, we use the global monthly precipitation data (PREC) created by Chen et al. (2002) instead of NCEP/NCAR precipitation data, because previous studies have shown that the NCEP/NCAR reanalysis precipitation exhibits some deficiencies and that this field does not compare as well with observations as other reanalysis fields, such as heights, winds, and temperatures that are assimilated directly into the model (Poccard et al. 2000; Janowiak et al. 1998). Monthly precipitation data have been selected from the global precipitation reconstruction data (PREC) estimates on a 0.5° × 0.5° latitude/longitude grid over the period 1948–2005 in Sudan, which has in total 890 grids. The PREC analyses are derived from gauge observations from over 17,000 stations collected in the Global Historical Climatology Network (GHCN), version 2, and the Climate Anomaly Monitoring System (CAMS) datasets. The global analyses are defined by interpolation of gauge observations over land (PREC/L) and by using empirical orthogonal functions (EOF) for interpolation and subsequent reconstruction of historical observations over ocean (PREC/O) (Chen et al. 2002). Shi et al. (2002a, b, 2004) analyzed the global land precipitation dataset (PREC/L) and found that this dataset was accurate enough for describing the change of large scale precipitation, and there was a good relationship between the global precipitation database (PREC) and observed Western Africa monsoon precipitation during 1948–2001. In the Northern Hemisphere, the tropical area to the south of 25°N, precipitation showed a negative trend in all seasons. For comparison, long-term mean monthly precipitation dataset from 39 observed rain gauge stations from GHCH Version 2 were also used in this study. The location of Sudan and observed and reconstructed grid points can be referred to Fig. 1 and Table 1. Figure 1 shows that the PREC data has, in general, a good agreement with observed data. In the rest of the paper detailed analysis is only done for the PREC data, because the observed data has too few stations and only 12 long-term mean monthly values are available.

Atmospheric data from NCEP/NCAR-I reanalysis (R-1) over the period 1948–2005 are available on a 2.5° × 2.5° latitude/longitude grid. The water vapor flux of the whole Ps layer (surface pressure) = 300 hPa was analyzed in this study. In the actual atmosphere, the moisture is very low over 300 hPa, so $p = 300 \text{ hPa}$ will be used in the calculation.

3 Methods

The Mann–Kendall trend test (MK; Mann 1945; WMO 1966; Kendall 1975; Sneyers 1990) is widely used in the literature to analyze trends in the climate data. The MK test has different variants. In this study two different versions are used for two different purposes. The procedure of the first version as demonstrated by Zhang et al. (2010a) starts by simply comparing the most recent data with earlier values. A score of $+1$ is awarded if the most recent value is larger, or a score of $-1$ is awarded if it is smaller. The total score for the time-series data is the Mann–Kendall statistic, $Z$, which is then compared to a critical value, $Z_{1–α/2}$ (where $α$ is significance level, and $Z_{1–α/2}$ is the Z value found in the standard normal distribution table), to test whether the trend in the data is significantly increasing ($Z > Z_{1–α/2}$), significantly decreasing ($Z < −Z_{1–α/2}$) or if no trend ($−Z_{1–α/2} < Z < Z_{1–α/2}$). This procedure is used to produce results as showed in Figs. 4 and 7, and in Table 2.

In contrast to the traditional MK test which calculates above statistic variables only once for the whole sample, MK method can also be used to test an assumption regarding the beginning of the development of a trend within a sample, i.e., a changing point in the time series (Zhang et al. 2009). Following the procedure as shown by Gerstengarbe and Werner (1999) who used the method to test an assumption about the beginning of the development of trend within a sample $(x_1, x_2, ..., x_n)$ of the random variable $X$, the corresponding rank series for the so-called retrograde rows are similarly obtained for the retrograde sample $(x_{n}, x_{n-1}, ..., x_1)$. Based on the rank series $r$ of the progressive and retrograde rows of this sample, the statistic variables, $Z_1$ and $Z_2$ will be calculated for the progressive and retrograde samples, respectively. The $Z_1$ and $Z_2$ values calculated with progressive and retrograde series are named UF and UB, respectively, in this paper. The intersection point of the two lines, UF and UB gives the point in time of the beginning of a developing trend within the time series. This method is used to produce results as showed in Fig. 5b.

As a complementary method of trend analysis, simple linear regression was also used in this paper for long-term linear trend test. The simple linear regression method is a parametric $T$-test method, which consists of two steps, fitting a linear simple regression equation with the time $t$ as independent variable and the hydrological variable (i.e.
precipitation and moisture flux in this study) \( Y \) as dependent variable; testing the statistical significance of the slope of the regression equation by the \( t \)-test (Xu 2001; Zhang et al. 2010a). The parametric \( T \)-test requires the data to be tested is normally distributed. The normality of the data series is first tested in the study by applying the Kolmogorov–Smirnov test. The \( T \)-test method is used to produce the results as showed in Figs. 5a and 9.

It is evident that the power of the test to detect the existing trend is significantly influenced by, among others, serial correlation in the time series and sample size or record length. The larger the sample size, the more powerful the test. Therefore, the assessment results of time series with different record length should not be put together to infer a general trend tendency (Yue and Wang 2002a, b). “Pre-whitening” is one of the methods used to prevent false indication of trend, where autocorrelation is removed from the data by assuming a certain correlation model, usually a Markovian one (e.g., Von Storch 1995).

The zonal moisture transport flux \( (Q_u) \) and meridional moisture transport flux \( (Q_v) \) were calculated based on the following equations (Cadet and Nnoli 1987; Zhou et al. 1998):

\[
Q_u(x, y, t) = \frac{1}{g} \int_{p_s}^p q(x, y, p, t)u(x, y, p, t)dp
\]

\[
Q_v(x, y, t) = \frac{1}{g} \int_{p_s}^p q(x, y, p, t)v(x, y, p, t)dp
\]
where \( u \) and \( v \) are the zonal and meridional components of the wind field, respectively; \( q \) is the specific humidity; \( p_s \) is the surface pressure; \( p \) is atmospheric top pressure; and \( g \) is the acceleration of the gravity (e.g., Cadet and Nnoli 1987; Miao et al. 2005; Zhang et al. 2010b).

### Table 1 List of rain gauge stations used in this study in Sudan

<table>
<thead>
<tr>
<th>Station No.</th>
<th>Name</th>
<th>Latitude (N)</th>
<th>Longitude (E)</th>
<th>Annual mean precipitation (mm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>62600</td>
<td>Wadi Halfa</td>
<td>21 49</td>
<td>31 21</td>
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</tr>
<tr>
<td>62615</td>
<td>Halaib</td>
<td>22 13</td>
<td>36 39</td>
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<tr>
<td>62620</td>
<td>Station No. 6</td>
<td>20 45</td>
<td>32 33</td>
<td>5.9</td>
</tr>
<tr>
<td>62635</td>
<td>Arbaat</td>
<td>19 50</td>
<td>36 58</td>
<td>35.6</td>
</tr>
<tr>
<td>62640</td>
<td>Abu Hamed</td>
<td>19 32</td>
<td>33 20</td>
<td>12.6</td>
</tr>
<tr>
<td>62641</td>
<td>Port Sudan</td>
<td>19 35</td>
<td>37 13</td>
<td>76.1</td>
</tr>
<tr>
<td>62650</td>
<td>Dongola</td>
<td>19 10</td>
<td>30 29</td>
<td>12.3</td>
</tr>
<tr>
<td>62660</td>
<td>Karima</td>
<td>18 33</td>
<td>31 51</td>
<td>20.7</td>
</tr>
<tr>
<td>62675</td>
<td>Aqiq</td>
<td>18 14</td>
<td>38 11</td>
<td>124.9</td>
</tr>
<tr>
<td>62680</td>
<td>Atbara</td>
<td>17 42</td>
<td>33 58</td>
<td>59.9</td>
</tr>
<tr>
<td>62682</td>
<td>Hudeiba</td>
<td>17 34</td>
<td>33 56</td>
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<tr>
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<td>Khartoum</td>
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<td>32 33</td>
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<tr>
<td>62722</td>
<td>Aroma</td>
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<tr>
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<td>Shambat Obs.</td>
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<td>32 32</td>
<td>127.5</td>
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<tr>
<td>62730</td>
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<td>36 24</td>
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<td>Halfa El Gedida</td>
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<td>35 36</td>
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<tr>
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<td>Showak</td>
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<td>35 51</td>
<td>501.9</td>
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<tr>
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<td>Gedaref</td>
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<td>35 24</td>
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<tr>
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<td>El Fasher</td>
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<td>30 14</td>
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<td>30 14</td>
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<tr>
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<td>En Nahud</td>
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<tr>
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<td>Nyala</td>
<td>12 3</td>
<td>24 53</td>
<td>398.3</td>
</tr>
<tr>
<td>62795</td>
<td>Abu Na’ama</td>
<td>12 44</td>
<td>34 8</td>
<td>556</td>
</tr>
<tr>
<td>62801</td>
<td>Renk</td>
<td>11 45</td>
<td>32 47</td>
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</tr>
<tr>
<td>62803</td>
<td>Rashad</td>
<td>11 52</td>
<td>31 3</td>
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</tr>
<tr>
<td>62805</td>
<td>Damazine</td>
<td>11 47</td>
<td>34 23</td>
<td>712.9</td>
</tr>
<tr>
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<td>Banabusa</td>
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<td>27 40</td>
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</tr>
<tr>
<td>62810</td>
<td>Kadugli</td>
<td>11 0</td>
<td>29 43</td>
<td>633.1</td>
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<tr>
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<td>Malakal</td>
<td>9 33</td>
<td>31 39</td>
<td>731.6</td>
</tr>
<tr>
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<td>Raga</td>
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<td>25 41</td>
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<td>Wau</td>
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<td>28 10</td>
<td>1074.5</td>
</tr>
<tr>
<td>62900</td>
<td>Rumbek</td>
<td>6 48</td>
<td>29 42</td>
<td>847.7</td>
</tr>
<tr>
<td>62941</td>
<td>Juba</td>
<td>4 52</td>
<td>31 36</td>
<td>972.7</td>
</tr>
</tbody>
</table>

### 4 Results

#### 4.1 Characteristics of annual precipitation

Before spatial and temporal variabilities of the annual precipitation data of PREC are examined, the standardized PREC data and the Sahel precipitation index are compared first. Results of comparison, shown in Fig. 2 for the Sahel area (10°–20°N, 20°W–10°E), reveal that the PREC data has a good agreement with the Sahel precipitation index in the study region. Spatial distributions of mean annual precipitation revealed by the PREC data are then compared with the interpolated observed data of 39 stations for the period 1961–1990 and the comparison is shown in Fig. 3. It can be seen from this figure that the average annual precipitation varies greatly in Sudan and it varies from almost nil in the north to more than 1000 mm in the Southwest. From the spatial angle, the distribution shows that rainfall is decreasing from south to north. The two-third of the whole country is controlled by arid and semi-arid climate with annual precipitation of less than 600 mm. The observed average annual precipitation between 1961 and 1990 also shows similar spatial patterns in Sudan (Fig. 3b). In general there is a good agreement between the two data sets. However, due to the limited number of stations some details of the spatial variability of mean annual precipitation as revealed in Fig. 3a are lost in b, especially in the rainfall rich region of South Sudan. Figure 3c estimates the magnitude and percentage of overestimations of PREC precipitation compared to the observed precipitation which provides important information for engineering designs. The results show that PREC precipitation are higher than observed precipitation, the amount of differences increases from north to south, however, the percentage of the increases is approximately 10% in the whole Sudan, except in the southeast which PREC is 17% higher than the observed values.

The trend analysis showed that the annual precipitation decreased in Sudan during 1948–2005 and the significant decreasing trend was found in central Sudan (Fig. 4). When the annual average precipitation over the whole country is concerned, a significant decreasing trend also exists during 1948–2005 (Fig. 5a). Furthermore, a study of year to year changes in the areal average annual precipitation by using the rank-based Mann–Kendall method (Gerstengarbe and Werner 1999) showed that a change point was found in around 1969 in the process (Fig. 5b). Similar results were obtained by Hulme (1990) when he reported that the depletion has been most severe in semi-arid Central Sudan. Other researchers, e.g., Osman and Shamseldin (2002), also found that the areal annual averaged rainfall values decreased markedly since the 1960s, and the drought in 1970s produces a large number of impacts that affects...
Sudan’s social, environmental, and economical standard of living with reduced crop, reduced water levels, increased livestock and wildlife death rates and damage to wildlife and fish habitat.

4.2 Monthly precipitation during the rainy season

As mentioned above, from January to March, Sudan is under the influence of dry north-easterlies. There is practically no rainfall countrywide except for a small area in northeastern Sudan (Fig. 1). The length of the rainy season decreases from about 9 months in the south to as little as 3 months on the southern fringes of the desert in the north, as revealed by Fig. 1. This study defines the rainy season in Sudan as the period between March and November.

The spatial distribution of average monthly precipitation for the period 1948–2005 in the rainy season is shown in Fig. 6. It is seen that regional scale precipitation starts in March and April, when the rainy season monsoons hit the southwestern area. In March, the moist southwesterlies reach southern Sudan bringing precipitation of an average magnitude of 70 mm to southern areas (Fig. 6a). In May, rains blow northwards, and southern Sudan gets more rainfall (Fig. 6c). By August, the rain belt extends to its usual northern limits and the precipitation amount reaches the maximum, causing southwest and southeast areas to receive an average of 280 mm of precipitation in a month (Fig. 6f). The rainy season in North Sudan onsets in June and extends to July, August, and September, ending eventually in October, and in November the monsoons retreat gradually from the country.

The temporal trends of the selected monthly precipitation in the rainy season and the annual totals are analyzed by using the MK test method (Table 2). The Z values indicate that significant decreasing trends exist in the main rainy months (July, August and September) with the steepest decrease found in August which is also the month receiving the maximum amount of rainfall in the country. Trilbach and Hulme (1984) also reported that heavy rains in the rain season declined more did than median and light rains. Table 2 also shows that decreasing trends exist in most months except October and November; consequently, the areal average annual precipitation is also decreasing significantly. The change point was found between 1967 and 1970 for the rainy season months. Previous studies have also noted that summer precipitation in the sub-Saharan region and Northeast China showed significant decreases around the year 1968, and precipitation over the Sahel region decreased sharply since then with drought persisting throughout the 1970s and 1980s (Lamb 1982, 1983; Quan et al. 2003).

For illustrative purposes, spatial variations of the MK trend for May–October are shown in Fig. 7. It is seen that in May an increasing trend in monthly rainfall is found in north Sudan while an insignificant decreasing trend occurs in most part of Sudan. In June and July, the country is dominated by a decreasing trend in monthly rainfall, but this trend is mostly insignificant except in one or two small regions which were marked as shadow areas. In August, however, a large part of central Sudan is dominated by significant decreasing trends, and in September a similar pattern as in July is found. There exists no significant temporal changing trend in October in Sudan except in the north-eastern corner where a significant increasing trend is found. From this picture, we can find that the rainfall decreases in the central Sudan during the rainy season.

4.3 Transition of precipitation and moisture flux anomalies

Sudan’s climate ranges from tropical to continental, while most parts of southern Sudan experience a monsoon

<table>
<thead>
<tr>
<th>Month</th>
<th>Mar</th>
<th>Apr</th>
<th>May</th>
<th>Jun</th>
<th>Jul</th>
<th>Aug</th>
<th>Sep</th>
<th>Oct</th>
<th>Nov</th>
<th>Annual</th>
</tr>
</thead>
<tbody>
<tr>
<td>MK value</td>
<td>-0.15</td>
<td>-0.01</td>
<td>-0.66</td>
<td>-1.15</td>
<td>-2.03</td>
<td>-3.44</td>
<td>-2.28</td>
<td>0.26</td>
<td>0.61</td>
<td>-3.38</td>
</tr>
</tbody>
</table>

* Statistically significant trends at the 0.05 levels, – None abrupt change point
climate. The amounts of precipitation and the length of rainy season depend on which of the following two air flows predominates: the dry northeasterly winds from the Arabian Peninsula or moist southwesterly winds from the Congo River basin. In the rainy season, the moist southwesterlies reach southern Sudan and bring more precipitation. By July the moist air will reach Khartoum, Sudan’s capital located in the central part, and in August it extends to its usual northern limits around Abu Hamad in the northern part. The flow becomes weaker as it spreads north. In September the dry northeasterlies begin to strengthen and to push south and by the end of December they cover the entire country.

It is therefore important to study the relationship between the African monsoon and the precipitation of Sudan and investigate how the changes in the African

Fig. 3 The average of annual mean precipitation based on Precipitation REConstrucion (PREC) (a), observed precipitation data (b) during 1961–1990, and PREC minus the observed precipitation (c) in Sudan. The dots in b and c stand for observation stations.
monsoon have influenced the amount of precipitation during the rainy season. It is also desirable to explore how the African monsoon has changed and whether this change is one of the main causes for the rainfall decrease and the abrupt change in annual and monthly precipitation found around 1968. Therefore, we analyzed the variability of atmospheric moisture content and moisture flux in the past decades and the average moisture flux anomalies to explain the transition of precipitation in Sudan. From Fig. 8, it is found that the whole layer of moisture content significantly has decreased in central Sudan in summer since the late 1960s which might be related to the decline in precipitation in central Sudan. The spatial distribution of temporal linear trend of the atmospheric moisture flux in 1948–2005 in Africa is shown in Fig. 9. It can be seen that the whole layer of moisture flux in summer (JJA) during 1948–2005 decreased significantly in the central part of Africa including Sudan. The significant negative trends of zonal and meridional moisture flux found in Sudan indicated that there was a weak western and southern moisture flux, respectively. Figure 10 shows the spatial distribution of summer moisture flux anomalies during 1970–2005, when compared to that during 1948–1969. It is seen that there was an obvious northeastern moisture flux in Sudan. The northeasterly wind tendency in Sudan indicated that there was a weak summer monsoon since the late 1960s, and the weak summer monsoon limited the northward propagation of moisture to North Sudan. As a result, precipitation decreases significantly, especially in central Sudan. This observation is in line with changes in precipitation in Sudan, showing a close agreement between moisture flux alterations and changes in precipitation. The changes in African monsoon in the late 1960s make the circulation and precipitation change. The decades of the 1970s and 1980s saw the southwesterlies frequently fail, with disastrous results for the Sudanese people and economy.

5 Discussions

The decrease in precipitation in rainy season has led to more severe and longer-lasting droughts in Sudan. Increasing drought frequency has the potential to affect land-based natural and managed ecosystems, coastal systems, and both freshwater quality and quantity (Alvi 1994; Janowiak 1988; Nicholson et al. 2000). The current study
reveals the annual and monthly changing characteristics of precipitation during 1948–2005 in Sudan. The annual precipitation has a significant decreasing trend in central Sudan, which is particularly caused by the decrease in the amount of rains in July, August and September. The decreasing trend in precipitation in Sudan, particularly after mid-1960s, is in agreement with regional precipitation changes as reported in earlier studies. For example, Eltahir (1988) and Long et al. (2000) found the annual rainfall amounts in the entire region of central and western Africa decreased since the late 1960s. The mechanisms of droughts in Sudan are complex and many studies have been made to examine the causes of droughts. However, the cause of the Sahel drought remains elusive (Nicholson

**Fig. 6** The monthly average precipitation in rainy season in Sudan averaged over the period 1948–2005.
Fig. 7 Spatial distribution of the MK trend for the selected months. The values of contour lines depict the test statistics $Z$ values of MK test. Solid lines indicate increasing regions and dashed lines indicate decreasing regions, and gray regions indicate the trend is statistically significant at the 5% significance level.
The objective here is to provide a complementary view to the previous studies by examining the relationship of moisture flux and precipitation trend in the region. Results confirmed that there is a relationship between moisture flux and precipitation variations in Sudan. However, it is by no means to say that the moisture flux is the only factor or more important factor than other factors that have been studied by other authors.

As a whole, the persistent drought over the sub-Saharan Africa (the Sahel region) is accompanied by the weakening summer monsoon and the significant decrease in precipitation might be associated with the weakening monsoons. The northeasterly wind tendency in the Sudan limits the northward propagation of moisture flux to North Sudan. Thus, the precipitation decreases significantly. Quan et al. (2003) found there are corresponding changes in the atmospheric circulation that are associated with the inter-decadal changes in summer precipitation over Asia and Africa. Long et al. (2000) also found that the variability of moisture flux is highly related to precipitation in Sudan. Therefore, the weakening African Summer Monsoon caused decreasing northward propagation of moisture flux to North Africa in general and in Sudan in particular. Fontaine et al. (2003) analyzed the atmospheric water and moisture fluxes in West African Monsoon and found that observed Sudan-Sahel rainfall and, in wet (dry) situations, with a clear dominance of westerly (easterly) anomalies in the moisture flux south of 15°N. Thus, the persistent drought over the sub-Saharan Africa (the Sahel region) might be accompanied by the weakened moisture flux in the region.

6 Conclusions

This study analyzed variations in precipitation and the whole layer of moisture flux during 1948–2005 in Sudan with the aim of exploring changes in precipitation and associated atmospheric circulation for the occurrence of precipitation transition in Sudan. Some interesting conclusions are obtained as follows:

1) The annual average precipitation varies greatly in Sudan from almost nil in the north to about 1500 mm in the extreme Southwest. August is the month with the highest precipitation rates in Sudan; however, the biggest decreasing trend also occurred in this month.

2) Precipitation decreases almost in all months in the rainy season and significantly decreasing trends can be found in rainy season and annually in the central part of Sudan. In August, a large part of central Sudan is dominated by significant deceasing trends.

3) The whole layer of moisture content significantly has decreased in central Sudan in summer since the late 1960s which might be one of the causes of the decline in precipitation in central Sudan. Abrupt changes in monthly and annual precipitation have occurred in the late of 1960s. The moisture flux over Sudan tends to be decreasing after these abrupt changes.

The existing clear relationship between moisture flux and precipitation variations in Sudan means that the changes in moisture flux is one of the main reasons for the spatial and temporal variation of precipitation in the region. This study provides a complementary view to the previous studies that attempted to explain the Sahel persistent drought.
drought in terms of general circulation features, sea surface temperatures, land surface feedback mechanisms, and teleconnected, spatiotemporal rainfall patterns.

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