IUFRO 9
RESEARCH SERIES

ENVIRONMENTAL CHANGE AND GEOMORPHIC HAZARDS IN FORESTS

EDITED BY R.C. SIDLE

IUFRO

CABI *Publishing*

IUFRO Research Series

The International Union of Forest Research Organizations (IUFRO), with its 14,000 scientists from 700 member institutions in over 100 countries, is organized into nearly 300 research units that annually hold approximately 80 conferences, workshops and other meetings. The individual papers, proceedings and other material arising from these units and meetings are often published but in a wide array of different journals and other publications. The object of the IUFRO Research Series is to offer a single, uniform outlet for high quality publications arising from major IUFRO meetings and other products of IUFRO's research units.

The editing, publishing and dissemination experience of CABI Publishing and the huge spread of scientific endeavours of IUFRO combine here to make information widely available that is of value to policy makers, resource managers, peer scientists and educators. The Board of IUFRO forms the Editorial Advisory Board for the series and provides the monitoring and uniformity that such a high quality series requires in addition to the editorial work of the conference organizers.

While adding a new body of information to the plethora currently dealing with forestry and related resources, this series seeks to provide a single, uniform forum and style that all forest scientists will turn to first as an outlet for their conference material and other products, and that the users of information will also see as a reliable and reputable source.

Although the official languages of IUFRO include English, French, German and Spanish, the majority of modern scientific papers are published in English. In this series, all books will be published in English as the main language, allowing papers occasionally to be in other languages. Guidelines for submitting and publishing material in this series are available from the Publisher, Books and Reference Works, CABI Publishing, CAB International, Wallingford, Oxon OX10 8DE, UK, and the IUFRO Secretariat, c/o Federal Forest Research Centre, Seckendorff-Gudent-Weg 8, A-1131, Vienna, Austria.

IUFRO Research Series

Titles available:

1. *Forest Dynamics in Heavily Polluted Regions*
   Edited by J.L. Innes and J. Oleksyn
2. *Forest History: International Studies on Socioeconomic and Forest Ecosystem Change*
   Edited by M. Agnoletti and S. Anderson
3. *Methods and Approaches in Forest History*
   Edited by M. Agnoletti and S. Anderson
4. *Air Pollution and the Forests of Developing and Rapidly Industrializing Countries*
   Edited by J.L. Innes and A.H. Haron
5. *Forests in Sustainable Mountain Development: a State of Knowledge Report for 2000*
   Edited by M. Price and N. Butt
6. *Forests and Landscapes: Linking Ecology, Sustainability and Aesthetics*
   Edited by S.R.J. Sheppard and H.W. Harshaw
7. *Criteria and Indicators for Sustainable Forest Management*
   Edited by J. Raison, A. Brown and D. Flinn
8. *The Impact of Carbon Dioxide and Other Greenhouse Gases on Forest Ecosystems*
   Edited by D.F. Karnosky, R. Ceulemans, G.E. Scarascia-Mugnozza and J.L. Innes
9. *Environmental Change and Geomorphic Hazards in Forests.*
   R.C. Sidle

Environmental Changes and Geomorphic Hazards in Forests

Report No. 4 of the IUFRO Task Force on Environmental Change

Edited by

Roy C. Sidle
Department of Geography
National University of Singapore
Singapore

CABI *Publishing*
in association with
The International Union of Forest Research Organizations (IUFRO)

CABI *Publishing* is a division of CAB *International*

CABI Publishing
CAB International
Wallingford
Oxon OX10 8DE
UK

Tel: +44 (0)1491 832111
Fax: +44 (0)1491 833508
E-mail: cabi@cabi.org
Web site: www.cabi-publishing.org

CABI Publishing
10 E 40th Street
Suite 3203
New York, NY 10016
USA

Tel: +1 212 481 7018
Fax: +1 212 686 7993
E-mail: cabi-nao@cabi.org

A catalogue record for this book is available from the British Library, London, UK.

Library of Congress Cataloging-in-Publication Data

Environmental change and geomorphic hazards in forests / edited by Roy C. Sidle
p. cm. -- (IUFRO research series ; v. 9)
Include bibliographical references.
ISBN 0-85199-598-5
1. Climatic changes. 2. Forest ecology. 3. Forests and forestry--Environmental aspects. 4. Hazardous geographic environments. I. Sidle, Roy C. II. International Union of Forestry Research Organizations. III. IUFRO research series ; 9.

SD390.7.C55 E58 2002
634.9'61--dc21

2002025903

ISBN 0 85199 598 5

Printed and bound in the UK by Biddles Ltd, Guildford and King's Lynn, from copy supplied by the editor.

Contents

Contributors

H.R. Beckedahl, Faculty of Science and Agriculture, Department of Geography, University of Natal, Pietermaritzburg 3209, South Africa, HBeck@nu.ac.za

C.R. Burn, Department of Geography and Environmental Studies, Carleton University, B349 Loeb Building 1125 Colonel By Drive, Ottawa, Ontario, Canada, KlS 5B6, crburn@ccs.carleton.ca

D.H. Burn, Department of Civil Engineering, University of Watérloo, Waterloo, Ontario, Canada N2L 3G1, dhburn@sunburn.uwaterloo.ca

Q. Cai, Institute of Geographical Sciences and Natural Resources, Chinese Academy of Sciences 917, Datum Road, Anwai, Beijing 100101, People's Rep. of China, caiqg@mx.cei.gov.cn

M. Chigira, Disaster Prevention Research Institute, Kyoto University, Gokasho, Uji 611-0011, Japan, chigira@slope.dpri.kyoto-u.ac.jp

A.S. Dhakal, Department of Forest Resources Management, University of British Columbia, 2424 Main Mall, Vancouver, British Columbia, Canada V6T 1Z4, dhakal@@interchange.ubc.ca

W. Haeberli, Department of Geography, University of Zurich, Winterthurerstrasse 190 CH–8057, Zurich, Switzerland, haebari@geo.unizh.ch

L.R. Leung, Pacific Northwest National Laboratory, PO Box 999, Richland, Washington, 99352, USA, ruby.leung@pnl.gov

K.F.A. Lo, Department of Natural Resources, Chinese Culture University, Hwa Kang, Yang Ming Shan, Taipei 111, Taiwan, ufb0043@ms5.hinet.net

G. Roldolfi, Department of Soil Science and Plant Nutrition, University of Florence, Piazzale delle Cascine 15, I–50144, Firenze, Italy, rodolfi@iges.fi.cnr.it

P.G. Sanderson, University of Notre Dame, Australia, School of Environmental Management, Freemantle, Western Australia 6959, psanderson@nd.edu.au

R.C. Sidle, Department of Geography, National University of Singapore, AS2 1 Arts Link, Republic of Singapore 117570, geosrc@nus.edu.sg. *New address:* Slope Conservation Section, Geodisasters Section, Disaster Prevention Research Institute, Kyoto University, Gokasho, Uji, Kyoto 611-0011, Japan

D. Taylor, Department of Geography, Trinity College, University of Dublin, Museum Building, Dublin 2, Republic of Ireland, taylord@tcd.ie

M.S. Wigmosta, Pacific Northwest National Laboratory, PO Box 999, Richland, Washington, 99352 USA

C. Zancih, Department of Agronomy and Land Management, University of Florence, Piazzale delle Cascine 15, I–50144, Firenze, Italy, camillo.zanchi@unifi.it

Foreword

The Task Force on Environmental Change was established in 1995. It's brief is to provide state-of-the-art summaries on various aspects of environmental change as it relates to forested areas. This is the fourth report of the Task Force to be published by CABI Publishing, and covers the important area of geohazards. As we begin to acknowledge more fully the many different values of forests, one that has received insufficient attention is the role of forests in mitigating natural hazards.

The protection forests of the European Alps have long been recognized as playing an important function in the protection of human infrastructure from avalanches, debris torrents and other geohazards. However, the roles of other types of forests have often been underestimated. For example, mangrove forests have a critical role in the protection of low-lying coastal regions from flooding and erosion. Mangroves are also critical as nurseries for many fish species of economic importance. Sea-level rise present a potential threat to such forests.

The important role that forests play in the mitigation of geohazards can be adversely influenced by forestry operations. Clear cuts, road-building and other forestry operations can all influence slope hydrology in such a way that the protective function of the forest cover is reduced. This is becoming an increasingly important issue in some parts of the world, to the extent that forestry operations have been severely curtailed in some areas. A better understanding of these interactions is very important if we are to mitigate the effects of forestry.

This report represents the state-of-the-art of our knowledge of most types of geohazards in forested areas. Scientists from around the world have contributed their expertise and I am delighted to how this cooperation between scientists has resulted in another excellent publication. I would also like to thank the Task Force coordinator, John Innes, for his continued efforts with these state-of-the-art reports.

Risto Seppälä
President, IUFRO

Acknowledgements

Gratitude is extended to the IUFRO Editorial Board for their support of this research series on environmental change. The Editor especially thanks John Innes, Coordinator of the IUFRO Task Force on Environmental Change, for his encouragement and support of this volume. The focus and clarity of the technical material presented in this book benefited significantly from numerous external reviews. Significant contributions from the following reviewers are gratefully acknowledged:

- Peter Hairsine, CSIRO Land & Water, Canberra, Australia
- John Innes, Department of Forest Resources Management, University of British Columbia, Canada
- Xi Xi Lu, Department of Geography, The National University of Singapore
- Hideaki Marui, Research Institute for Hazards in Snowy Areas, Niigata University, Japan
- Yukinori Matsukura, Institute of Geoscience, University of Tsukuba, Japan
- Paul Pilon, Meteorological Service of Canada, Ontario Region, Canada
- Dino Torri, CNR-IGES, Firenze, Italy
- Andrew Whitaker, Graduate School of Science & Technology, Niigata University, Japan
- Colin Woodroffe, School of Geosciences, University of Wollongong, Australia

Support from University of British Columbia and The National University of Singapore allowed the Editor to complete this volume. The later included funding from NUS research grant #R-109-000-031-112. The following people assisted in the formatting and technical details of the book: Karin Laursen, Junjiro Negishi, Anne Quah, Nabil Rohman, and Kwok Hoong Wu.

Chapter 1

Geomorphologic Hazards and Forest Environmental Change – an Introduction

Roy C. Sidle
Department of Geography, The National University of Singapore, Singapore

Overview of the Issues Affecting Geomorphic Hazards

We live in a world that is constantly changing, not merely from a perspective of climate and human activities, but also the more recent surge in information technology. Such developments affect how the public perceives geomorphic hazards. These are no longer alien concepts; however, as scientists, land managers, and planners it is important that we ensure that accurate, rational, and timely information and assessments are disseminated to the public (Sorensen, 2000). Neither apocalyptic nor 'business as usual' scenarios of climate change and forest land use are in the best public interest. Due to the highly interactive nature of geomorphic hazards related to feedbacks from climate change and land use, simplifications are often conveyed by the media that mask the complexity of ecosystem behaviour and portray inappropriate cause and effect relationships; some of these inappropriate generalizations have made their way into the scientific literature (e.g., Friedman and Friedman, 1988; Portela and Aguirre, 2000). What is needed is a systematic collection of scientific information, identification of information gaps, and new research information in key areas to squarely address the combined influences of climate change and changing forest land use on geomorphic hazards. This volume with contributions from five continents has made some strides in that direction. Herein, forested land is viewed from the broadest possible perspective, including rangeland, dry shrub land, and previously forested land that has been (or likely will be) converted to predominantly agricultural land. It is in this latter case where many of the most pressing and challenging global change issues reside (Byron and Arnold, 1999; Laurance, 1999; Fischer and Vasseur, 2000).

For some geohazards, such as surface erosion, desertification, and related land degradation, the impacts of anthropogenic change appear to outweigh effects of potential global warming. This is particularly true in developing nations where forest lands are rapidly being converted to agricultural plantations that require more roads and greater use of machinery and agricultural chemicals. It is also true in rapidly urbanizing regions, including areas receiving large influxes of tourists. Geomorphic hazards that more directly respond to temperature change will be most affected by global warming. These include rain-on-snow flood events, failures in weathered rock, mass wasting related to melting of glaciers and permafrost, snow avalanches, and coastal flooding and inundation as the result of sea-level rise. However, as is the case for most landslide, coastal and flood hazards, it is the combination of climate change and anthropogenic impacts that must be evaluated concurrently to predict the overall trajectory of geomorphic hazard response in a changing world.

Predictions of climate change are difficult at spatial scales that influence geomorphic hazards. The spatial resolution of climatic variables that influence geomorphic hazards varies from less than a km^2 to tens of km^2; global circulation models are unable to capture such resolution (e.g., Buma and Dehn, 1998). Likewise, predicting the trajectories of anthropogenic impacts is complicated. Land use is partly controlled by technological developments (e.g., changes in timber harvesting techniques and hillslope cultivation practices) and partly by socio-political issues (e.g., changing demographics, wealth distribution, and state-sponsored development). The technological 'advances' may allow new types of disturbances and ecosystem perturbations to occur in forested and formerly forested areas, thus exacerbating certain geomorphic hazards, such as landslides, desertification, and surface erosion. On the other hand, socio-economic factors typically expose low-income and subsistence inhabitants in developing and developed countries to more vulnerable situations. A stark example is the high vulnerability to flooding of dwellers in the extensive lowlands of Bangladesh (Kubo, 1993), as well as the increased potential vulnerability related to the effects of global warming on sea-level rise. Nevertheless, scenarios of likely climate change and anthropogenic impacts can be developed and assessed in regions that are at high risk for geomorphic hazards.

The Hazards

Many, but not all, of the geomorphic hazards that impact forests, rangelands, and steep land agriculture are discussed in this volume. The focus has been placed on hazards that impact large areas of forested or formerly forested terrain or present high risks to humans or natural resources. Most hazards are triggered or propagated by geomorphic thresholds, either extrinsic or intrinsic (Schumm, 1973). An extrinsic threshold is one that is exceeded by an external forcing variable, such as rainfall triggering a landslide (Sidle *et al.*, 1985). Climate change can also define or reset extrinsic thresholds, as in the case of severe erosion following a shift in environmental variables such as ground cover (e.g., Prosser and Soufi, 1998). Intrinsic geomorphic thresholds are achieved by long-term changes in the properties of earth surface materials (e.g., weathering as a precursor to rock fall); however, if alterations in materials are influenced by environmental change, the respective thresholds are considered to be at least partly extrinsic.

Chapters 2 and 3 of this volume cover the effects of climate change on flooding and river flow regimes in forested areas – both from a perspective of modelling snowmelt and rain-on-snow events, as well as based on recent historical records at a large regional scale (Canada). Chapters 4, 5, and 6 assess the more chronic but equally significant hazards related to surface erosion and land degradation, including desertification. These chapters adopt regional foci – Chapter 4 addresses surface erosion and land degradation in Asia and the tropical Pacific; Chapter 5 covers erosion and desertification in Mediterranean Europe; and Chapter 6 discusses erosion issues in the context of social developments in Africa. Chapter 7 focuses on the effects of climate change on weathering processes related to slow and long-term gravitational deformation of rocks and rock slides, while Chapter 8 specifically addresses the interaction of forest practices on landslides in the context of global change. The implications of climate change on permafrost and glacial hazards in high latitude forested sites are covered in Chapter 9. Chapter 10 evaluates the role and response of coastal tropical forests (particularly mangroves) to climate and environmental changes. The final chapter (Chapter 11) provides an overview of the emerging approaches for assessing the occurrence and impacts of geomorphic hazards in changing environments together with future research and application challenges.

Several geomorphic hazards are not addressed in individual chapters, either because of their regional specificity or because they do not typically or uniquely interact with forested or formerly forested ecosystems. Volcanic hazards are discussed in the context of the susceptibility of volcanic materials to surface erosion (Chapter 5), weathering (Chapter 7) and landslides (Chapter 8), as well as hazards on ice-covered volcanoes (Chapter 9). Likewise, snow avalanches are not specifically covered, but other hazards related to snow, such as flooding during snowmelt (Chapters 2 and 3), snowmelt-triggered landslides, debris flows and lahars (Chapters 8 and 9), and the interactions of snow cover and melt with weathering and permafrost processes (Chapters 7 and 9) are discussed. Earthquakes are discussed as they influence surface erosion (Chapters 4 and 5) and mass failures (Chapters 7, 8, and 9). Fire, while not a geomorphic hazard *per se*, is an important component of environmental change due to its frequent use in evolving land management. Also, the frequency and severity of wildfire can increase in scenarios of global warming due to changes in temperature, soil moisture, and humidity (Pinol *et al.*, 1998; Miller and Urban, 1999), as well as numbers of lightning strikes (Price and Rind, 1994). Thus, effects of fire are included in the chapters on surface erosion (Chapters 4 and 5), landslides and rock failures (Chapters 7 and 8), permafrost (Chapter 9) and tropical coastal forests (Chapter 10).

The regional examples included in this volume were selected to emphasize situations where global environmental change would have potentially large impacts on forest and rangeland ecosystems. For example, the flooding examples are drawn from temperate regions where snow is an important component of the water balance. Thermal conditioning of snowpacks that leads to peak runoff can be greatly influenced by climate change and by the management of forest cover. Surface erosion and land degradation scenarios were selected for three of the most vulnerable regions of the world: Asia, Mediterranean Europe, and Africa. This vulnerability ranges from the erodibility of soil materials (Chapter 4) to the stresses of changing land use (Chapter 5) to socio-economic impacts of degraded land on local populations (Chapter 6). The chapter on coastal forests (Chapter 10) focuses on tropical regions because of their potential to be impacted by sea-level rise and the ongoing effects of mangrove destruction and adjacent land conversion. The chapter on glacial and permafrost hazards (Chapter 9) is naturally confined to high latitudes, while the weathering and landslide chapters (Chapters 7 and 8) are more global in reference and application due to the widespread nature of these geomorphic hazards.

The Approaches

Various methodologies for assessing the effects of global change in forested environments on geomorphic hazards are presented in this volume. To some extent, the approaches selected are dictated by the nature of not only the hazard, but also the environmental context in which the hazard occurs. For example, glacial and permafrost hazards affect few people but future global warming may potentially impact large land areas; however, widespread soil erosion in Africa, while certainly a significant resource loss, has a direct impact on the livelihood and welfare of a major segment of the population. Thus, the emphasis in Chapter 9 is much more resource- and process-focused compared to the socio-economic focus of Chapter 6.

Historical records of extreme climatic events are frequently used in statistical analyses to estimate future trajectories of geomorphic and hydrologic response. Simple quantitative comparisons of historical sediment transport and erosion data provide compelling evidence of environmental changes (e.g., Chapters 4, 5, and 6). Trend analysis was applied to as many as 50 years of catchment discharge data (including peak flow, low flow, mean discharge, and flow variability) within three climatic regions of Canada (Chapter 3). Such statistical models are useful for detecting past trends, but caution must be exercised in applying such trends to forecasts of hydrologic response in changing climate conditions unless they are directly linked to climate change variables. Similar cautions must be exercised in applying extreme value statistics to peak and low flow analysis. Strong inferences from process studies can be applied to climate and anthropogenic change scenarios. While these are not direct measurements of environmental change, they provide insights into how geomorphic processes such as landslides, weathering, glaciation, and permafrost development would change under different conditions. Small-scale, controlled or conditional experiments can also elucidate aspects of environmental changes; examples presented in this volume include retrospective studies of bedrock weathering (Chapter 7) and plot experiments on the effect of surface cover changes on permafrost location (Chapter 9). Additionally, retrospective examinations of environmental change can be assessed through fossil, pollen, and stable isotope records as well as local knowledge of historical trends (Chapter 10). The use of various types of models to predict geomorphic hazard response to climate and anthropogenic changes holds promise, but the incorporation of such changes into models are currently relegated to scenario estimation (e.g., Ramakrishnan, 1998). Empirical models for soil erosion (e.g., the Universal Soil Loss Equation, see Chapter 6) can incorporate changes in the rainfall and runoff factor and the cover and management factor to approximate environmental change scenarios (Dissmeyer and Foster, 1980), however, it is doubtful that these will provide realistic estimates of probable erosion changes because of the inability of the model to capture episodic event-driven processes that often dominate surface erosion. The availability of regional data and practical knowledge of the application of the Universal Soil Loss Equation are its greatest assets (Chapter 6). Numerous combined empirical and process-based erosion models that are either distributed or semi-distributed have evolved. Such a model (AGNPS, Agricultural Non-Point Source Pollution model) is applied to assess erosion on converted hillslopes in Taiwan (Chapter 4). While sediment routing at the catchment scale is better characterized in such distributed models, they suffer from the same limitations of relying on uncertain future storm intensity information. Other more process-based distributed models have been developed for streamflow (DHSVM, Wigmosta *et al.*, 1994; see Chapter 1) and shallow landslides (dSLAM, Wu and Sidle, 1995; see Chapter 7). Advances in digital terrain analysis and remote sensing will undoubtedly enhance the application and availability of such complex models to broader geographic regions.

Perspective

While climate modellers struggle to develop yet more sophisticated global circulation models to predict temporal and spatial climatic response to anticipated increases in CO_2 and other aerosols, forest and range managers are coping with long-term decisions related to land use at very different spatial scales. The more detailed spatial scale at which land managers operate will not change, however, an increased awareness of potential regional changes in climate and their impact on geomorphic hazards is needed to advance our current ability to assess risk and vulnerability to floods, droughts, severe erosion and land degradation, landslides and rock failures, snow and ice hazards, and effects of rising sea levels in forested and formerly forested terrain. The chapters that follow in this volume present some important information, scenarios, and insights related to this overall objective.

References

Buma, J. and Dehn, M. (1998) A method for predicting the impact of climate change on slope stability. *Environmental Geology* 35, 190-196.

Byron, N. and Arnold, M. (1999) What futures for the people of the tropical forests? *World Development* 27, 789-805.

Dissmeyer, G.E. and Foster, G.R. (1980) A guide for predicting sheet and rill erosion on forest land. *USDA Forest Service Tech. Publ. SA-TP 11*, Southeast Area, Atlanta, Georgia, USA.

Fischer, A. and Vasseur, L. (2000) The crisis in shifting cultivation practices and the promise of agroforestry: a review of the Panamanian experience. *Biodiversity and Conservation* 9, 739-756.

Friedman, S.M. and Friedman, K.A. (1988) *Reporting on the Environment: A Handbook for Journalists.* U.N. Economic and Social Commission for Asia and the Pacific (ESCAP), Bangkok, Thailand.

Kubo, S. (1993) Geomorphological features of northwesten Bangladesh and some problems on flood mitigation. *GeoJournal* 31, 313-318.

Laurance, W.F. (1999) Reflections on the tropical deforestation crisis. *Biological Conservation* 91, 109-117.

Miller, C. and Urban, D.L. (1999) Forest pattern, fire, and climatic change in the Sierra Nevada. *Ecosystems* 2, 76-87.

Pinol, J., Terradas, J. and Lloret, F. (1998) Climate warming, wildfire hazard, and wildfire occurrence in coastal eastern Spain. *Climatic Change* 38, 345-357.

Portela, A.H. and Aguirre, B.E. (2000) Environmental degradation and vulnerability in Cuba. *Natural Hazards Review* 1, 171-179.

Price, C. and Rind, D. (1994) The impact of a 2×CO_2 climate on lightning-caused fires. *Journal of Climate* 7, 1484-1494.

Prosser, I.P. and Soufi, M. (1998) Controls on gully formation following forest clearing in a humid temperate environment. *Water Resources Research* 34, 3661-3671.

Ramakrishnan, P.S. (1998) Sustainable development, climate change and tropical rain forest landscape. *Climatic Change* 39, 583-600.

Schumm, S.A. (1973) Geomorphic thresholds and complex response of drainage systems. In: Morisawa, M. (ed.), *Fluvial Geomorphology*. SUNY, Binghamton, NY, Publ. Geomorphology, pp. 299-310.

Sidle, R.C., Pearce, A.J. and O'Loughlin, C.L. (1985) *Hillslope Stability and Land Use.* Water Resources Monograph, Vol. 11, American Geophysical Union, Washington, DC, 140 pp.

Sorensen, J.H. (2000) Hazard warning systems: review of 20 years of progress. *Natural Hazards Review* 1, 119-125.

Wigmosta, M.S., Vail, L.W. and Lettenmaier, D.P. (1994) A distributed hydrology-vegetation model for complex terrain. *Water Resources Research* 30, 1665-1679.

Wu, W. and Sidle, R.C. (1995) A distributed slope stability model for steep forested hillslopes. *Water Resources Research* 31, 2097-2110.

Chapter 2

Potential Impacts of Climate Change on Streamflow and Flooding in Snow-Dominated Forested Basins

Mark S. Wigmosta and L. Ruby Leung
Pacific Northwest National Laboratory, Richland, Washington, USA

Changes in climate resulting from increased atmospheric greenhouse gases may affect the frequency and magnitude of flooding in forested watersheds through changes in snowpack, soil moisture, and runoff production. The level of impact will vary among watersheds depending on climate regime and hydrologic characteristics of the catchments. Two forested watersheds in the Pacific Northwest, USA, the American River and Middle Fork Flathead River, were studied to examine the influence of long-term global warming on streamflow and flooding in snowmelt-dominated basins. These watersheds were selected to compare impacts resulting from changes in the maritime and continental climate regimes associated with the American and Middle Fork Flathead River basins, respectively. Output from a regional climate model was used to drive a distributed hydrologic model under present and future ($2\times CO_2$) climate conditions. For the future climate scenario more winter precipitation fell as rain instead of snow producing higher winter flows, a reduced snowpack, and decreased spring and summer flows in the American River. In addition, there was a large increase in the frequency and magnitude of winter flooding, primarily due to an increase in the number of rain-on-snow events. The change was much less dramatic in the Middle Fork Flathead River where flooding generally occurs during spring snowmelt. In this basin, the seasonal pattern of streamflow remains intact and the incidence of flooding was reduced for the future climate scenario. This study suggests that the impacts of climate change on streamflow and flooding in forested watersheds are highly region specific.

Introduction

The hydrological system of forested watersheds is potentially quite sensitive to changes in climate. Changes in precipitation and air temperature may impact the magnitude and timing of streamflow through changes in soil moisture, interception loss, evapotranspiration, and runoff production. Changes to the temporal pattern of precipitation, even when the mean remains constant, may change the frequency and intensity of flooding (IPCC, 1996). Snow-dominated catchments are particularly vulnerable to changes in climate. Relatively minor increases in air temperature may reduce snow accumulation, shift the snowmelt season from spring to winter, reduce summer flows, and produce more frequent rain-on-snow flooding in the autumn and winter (Lettenmaier and Gan, 1990; Leung and Wigmosta, 1999).

Changes in the seasonal pattern of runoff and the frequency of extreme runoff-producing events can significantly impact the geomorphological hazards in forested areas. Increases in mean and extreme precipitation will result in higher soil-water pore pressures and the increased incidence of rockfalls, landslides, and debris flows in areas prone to these hazards. This may in turn lead to increased sediment loading to channels. The situation may be exacerbated if more frequent summer droughts reduce forest cover through fire, increased pest infestation, and reduced growing season soil moisture.

Climate-induced changes in flood frequency and sediment loading will probably trigger changes in stream channel morphology. The locations of channel instability and sites of erosion and sedimentation may also change. Changes in morphology will be greater in alluvial channels than constrained channels at or near bedrock. Any changes in channel morphology, sediment concentrations, and stream temperature are likely to impact streamside habitat and the aquatic ecosystem.

Current climate models predict an increase in global mean precipitation of around 3–15% for an air temperature increase between 1.5–4.5°C (IPCC, 1996). Precipitation is expected to increase at higher latitudes, particularly in the winter. Details on the changes in regional precipitation are not clear, although the timing and patterns of precipitation are likely to change, with increases in some areas and decreases in others. There is also the possibility of more intense precipitation events in some regions (Gleick, 1999).

The effects of given changes in climate on streamflow vary considerably among watersheds depending on the climatic regime and the physical and biological characteristics of the catchments. Impacts to streamflow in rain-dominated watersheds will be largely propagated through changes in rainfall patterns and potential evapotranspiration. Increases in air temperature and net radiation will increase potential evapotranspiration in many areas, subject to associated changes in humidity and wind. The actual rate of evapotranspiration is constrained by soil moisture availability. In some regions an increase in summer rainfall, and therefore soil moisture, may have little impact on streamflow because of increased evapotranspiration, whereas an increase in winter rainfall is likely to produce a more noticeable change.

Increases in the frequency of high intensity rainfall will be felt more strongly in smaller basins, with thin, low permeability soils. Streamflow in these basins responds rapidly to short-duration, high intensity rainfall. Larger basins with good vegetative cover and deeper, more permeable soils will be more responsive to longer-duration rainfall under wet antecedent conditions and cumulative rain events.

Snow-dominated watersheds are particularly sensitive to potential changes in climate (Lettenmaier and Gan, 1990; Leung and Wigmosta, 1999). Higher air temperatures alone under greenhouse warming can result in more precipitation falling as rain rather than snow, which implies higher and more frequent winter floods, a reduced snowpack, and decreased spring and summer flows. More frequent rain-on-snow events are likely to increase wintertime flooding with or without an increase in the number of high rainfall events. The magnitude of these effects, which depends on the local changes in surface air temperature and precipitation for different existing climate (e.g., maritime versus continental) and hydrologic regimes, is expected to be highly variable at the regional level.

The likely impacts from climate change on streamflow and flooding in forested basins is highly dependent on location, climate, and the dominant forms of runoff production. We present a case study that focuses on two forested watersheds in the Pacific Northwest of the United States to examine the impacts of climate change on streamflow and flooding in snowmelt-dominated basins. Numerical experiments were performed to simulate the present and future regional climate conditions when the CO_2 concentration is double the present value. These simulations were used to drive a distributed hydrology model in the two watersheds to study potential changes in flow conditions.

Case Study Description

The two study watersheds are the American River watershed located on the east slope of the Cascade Range in Washington State, and the Middle Fork Flathead River watershed located on the west slope of the Northern Rockies in Montana. These watersheds were selected to compare the impacts resulting from potential changes in the maritime and continental climate regimes. In an earlier paper (Leung and Wigmosta, 1999) we discussed in detail the influence of potential climate change on seasonal streamflow in these two watersheds. This chapter provides a summary on changes in seasonal flow, but focuses mainly on changes in the frequency and magnitude of flood events. In what follows, we describe the watersheds, the climate scenarios, the hydrologic model used, and the impacts on streamflow and flood frequency.

Description of the watersheds

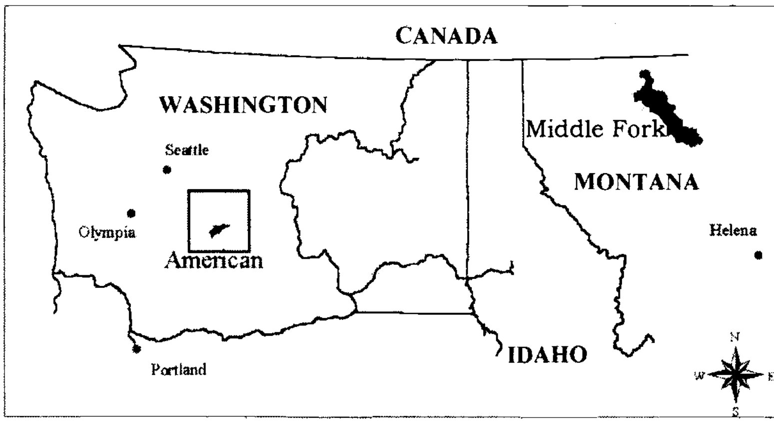


Fig. 2.1. Geographic locations of the American and Middle Fork Flathead River Basins.

Fig. 2.1 shows the geographical locations of the watersheds. The American River watershed covers an area of 204 km^2 and spans an elevation range between 849 m (at the channel outlet) and 2102 m (at the ridges). The American River is located in a coastal mountain range but receives on average only 1956 mm of precipitation annually at the Morse Lake SNOTEL site (elevation: 1646 m) because of rainshadowing by the Cascade mountains. In comparison, 2972 mm of precipitation falls at the Paradise meteorological station that is located at an elevation 86 m lower, but to the west on the upwind slope of the Cascades. Runoff is snowmelt dominated, typically with rain-on-snow events in the autumn and early winter, and radiation and temperature driven melt during the spring. There is a large natural variability in the amount and timing of streamflow that depends on the amount and phase of the precipitation during the cold season.

The Middle Fork Flathead River basin covers an area of 2916 km^2 and spans an elevation range between 959 m and 3047 m. The mean annual precipitation at the Badger Pass SNOTEL site (elevation: 2103 m) is about 1422 mm. Being located further inland, only about 60% of annual precipitation falls during the cold season at Badger Pass as compared to 80% at Morse Lake, reflecting the characteristics of continental as opposed to maritime climate. During winter, precipitation mostly falls in the form of snow because of the cold air temperature. Runoff is also snowmelt dominated, with a single peak that normally occurs between May and June.

Climate scenarios

Two regional climate simulations were used to study the impacts of climate change on mountain water resources. They include a 7-year control simulation of the present climate, and an 8-year simulation of the $2\times CO_2$ climate. The National Center for Atmospheric Research (NCAR) Community Climate Model (CCM3) (Kiehl *et al.*, 1996) was used to simulate the global circulation. This model was selected because it produces reasonable simulations of the observed large-scale conditions over the Pacific Northwest. The model was driven by the observed ocean conditions for 1989–95 in the control simulation, and the ocean conditions simulated by the Geophysical Fluid Dynamics Laboratory (GFDL) coupled ocean-atmosphere model in which CO_2 concentration increases at the rate of 1% per year until it reaches the level of 680 ppm or roughly double the present concentration of 340 ppm (Manabe *et al.*, 1991). According to the business-as-usual CO_2 emission scenario (e.g., IPCC, 1996), atmospheric CO_2 concentration will reach 680 ppm by around 2080.

Both the control and $2\times CO_2$ CCM3 simulations were used to drive the Pacific Northwest National Laboratory Regional Climate Model (RCM) (Leung and Ghan, 1995, 1998) at 90 km explicit spatial resolution. Data used to drive the RCM includes 12-hourly general circulation model (GCM) simulations of wind, air temperature, water vapour mixing ratio, and surface pressure applied at the RCM lateral boundaries. The RCM features a subgrid parameterization of orographic precipitation and land cover that divides each RCM grid cell into a limited number of surface elevation/vegetation bands. Physical processes such as cloud and precipitation, radiation, turbulent transfer, and surface physics are all calculated for each subgrid band. According to Leung and Ghan (1999a,b), under the $2\times CO_2$ conditions, there is an average warming of about 2°C and precipitation generally increases over the Pacific Northwest and decreases over California. The CCM3 and RCM simulations have been described in detail by Leung and Ghan (1999a,b).

Following the approach of Leung *et al.* (1996), RCM simulations for each surface elevation/vegetation band of each RCM grid cell were mapped to the higher-resolution (smaller) distributed hydrology model grid cells contained within the band. Data used to drive the Distributed Hydrology-Soil-Vegetation Model (DHSVM) include daily maximum and minimum air temperature, precipitation, relative humidity, wind, surface pressure, and downward solar radiation. The annual precipitation of the control simulation compares very well with observations in both watersheds, despite some errors in the seasonal distribution of precipitation. Leung and Ghan (1999a) showed that when the RCM was driven by the CCM3, there is a general cold bias during spring and a warm bias during summer. Errors in the seasonal distribution of precipitation and these air temperature biases can cause a shift in the simulated hydrographs under both the control and $2\times CO_2$ climate conditions. To simulate the greenhouse warming streamflow signal based on the existing hydrologic regimes, bias corrections for precipitation and surface air temperature were developed for each watershed based on the difference between the basin-averaged monthly precipitation and mean air temperature from observations and those from the control simulation. In each watershed, the same corrections were applied to both the control and 2×CO_2 climate scenarios. This method differs slightly from the approach described in Leung and Wigmosta (1999) where bias corrections were made for air temperature only, using the difference between regionally-averaged monthly mean air temperatures from observations and the control simulations.

Hydrologic model

The Distributed Hydrology-Soil-Vegetation Model (DHSVM) (Wigmosta *et al.*, 1994; Storck *et al.*, 1998; Wigmosta and Lettenmaier, 1994, 1999) was used to simulate hydrologic conditions in the watersheds. DHSVM provides an integrated representation of watershed processes at the spatial scale described by digital elevation model (DEM) data (100m horizontal resolution for the American River and 200m resolution for the larger Middle Fork Flathead River). It includes a multi-layer canopy model for evapotranspiration, energy balance models for canopy snow interception and ground snowpack processes, a multi-layer rooting zone model, and subsurface, surface, and channel flow modules. The modelled landscape is divided into computational grid cells centred on DEM elevation nodes. Digital elevation data are used to model topographic controls on absorbed shortwave radiation and down-slope water movement. At each time step, the model provides a simultaneous solution to the energy and water balance equations for every grid cell in the watershed. Each cell exchanges water with its adjacent neighbours, resulting in a three-dimensional redistribution of surface and subsurface water across the landscape.

Each grid cell is assigned a surface cover type and soil properties. These properties may vary spatially throughout the basin. Local meteorological conditions from the regional climate model were specified for each cell at a specified distance above the canopy. Solar radiation and wind speed are attenuated through each canopy layer based on cover density and Leaf Area Index (LAI), providing separate values for each canopy layer and the soil. Evapotranspiration from vegetation is modelled using a multi-canopy representation with each canopy layer partitioned into wet and dry areas. Snow accumulation and melt below the canopy or in the open are simulated using a two-layer energy and mass-balance model that explicitly incorporates the effects of topography and vegetation cover on the energy exchange at the snow surface.

The vertical movement of unsaturated soil moisture is simulated using a multi-layer soil rooting zone model. Discharge is calculated via Darcy's Law assuming a unit hydraulic gradient using the vertical unsaturated hydraulic conductivity of the soil (Wigmosta *et al.*, 1994). In each grid cell, percolation from the lowest rooting zone recharges the local water table. Every grid cell is allowed to exchange water with its adjacent neighbours as a function of local hydraulic conditions; thus, a given grid cell will receive water from up-gradient neighbours and discharge to adjacent down-gradient neighbours. If the computed water table in a grid cell is above the ground surface, the excess volume represents soil water exfiltration to the surface, which contributes to surface runoff. Rainfall or snowmelt occurring on saturated areas contributes directly to surface runoff.

DHSVM has been applied to the Middle Fork Flathead River when driven by observed (Wigmosta *et al.*, 1994) and RCM simulated (Leung *et al.*, 1996) meteorology. The simulated streamflow and snowpack compared very well with observations. In each watershed, the model has been calibrated and validated using observed meteorology, streamflow, and snow water equivalent (SWE) measurements for a period of 8 years between 1988–1995. DHSVM is applied at 100 m and 200 m horizontal resolution respectively in the American River and the Middle Fork Flathead River to best simulate the observed hydrologic conditions.

Case Study Results

Mean annual and mean monthly values were used to compare DHSVM simulated streamflow from the 7 year control run with the 8 year simulation under the $2\times CO_2$ climate. A threshold analysis was used to evaluate changes in flooding. The threshold for flooding in each basin was taken as the daily discharge equalled or exceeded by 1% of the daily flows in that basin's control run (1% daily control discharge threshold). The threshold discharge for the American River was 17 mm d^{-1}, slightly above the 2-year, 24-h flow of 16.4 mm d^{-1} (based on log-Pearson type III fit to 60 years of daily flow measurements). The threshold discharge for the Middle Fork Flathead River was 12.8 mm d^{-1}, corresponding to a return interval between 1.25 and 2 years (11.9 and 15.5 mm d^{-1}, respectively) based on 58 years of records. The number of days per year at or above this threshold was tabulated (~3.6 for control) along with the mean discharge and total volume per year. The same threshold value was used in the corresponding $2\times CO_2$ runs and changes in the frequency, mean discharge, and volume were evaluated. A similar approach was used to evaluate changes in high intensity precipitation events. The threshold for precipitation in each basin was taken as the total daily precipitation that was equalled or exceeded by 1% of the days in that basin's control run (1% daily control precipitation threshold).

Seasonal streamflow

Simulated annual precipitation increased by 7% and 8% under $2\times CO_2$ conditions in the American and Middle Fork Flathead watersheds, respectively (Table 2.1). Mean annual air temperature shows a warming of 2.1°C and 1.9°C in these two basins. The warming is stronger in the winter than the summer.

Table 2.1. Annual basin-mean conditions for control and $2\times CO_2$ simulations in the (a) American River, and (b) Middle Fork Flathead River watersheds.

Variables	Control simulation	$2\times CO_2$ simulation	Change
	(a) American River		
Precipitation (mm)	1,807	1,933	126 (7%)
Air temperature (°C)	3.4	5.5	2.1 (62%)
Streamflow (mm)	951	1,089	138 (15%)
ET (mm)	856	844	-12 (-1%)
Max. monthly SWE (mm)	612	216	-396 (-65%)
	(b) Middle Fork Flathead River		
Precipitation (mm)	1,342	1,453	111 (8%)
Air temperature (°C)	1.5	3.4	1.9 (127%)
Streamflow (mm)	700	763	63 (9%)
ET (mm)	550	598	48 (9%)
Max. monthly SWE (mm)	541	431	-110 (-20%)

Mean monthly hydrographs simulated by DHSVM for the present and $2\times CO_2$ conditions for the two watersheds are presented in Figs. 2.2 and 2.3. The change in the hydrologic conditions is rather drastic in the American River (Fig. 2.2). Although the peak discharge remains similar, there are large changes in both the annual total (Table 2.1) and the timing of streamflow. Streamflow in the December–March period under $2\times CO_2$ is nearly double that of the control simulation. This is both a result of the higher precipitation and warmer air temperature that causes a higher percentage of precipitation to fall in the form of rain rather than snow. The latter causes a more immediate contribution to the runoff. Because of the warmer air temperature, the basin maximum monthly snow water equivalent (SWE) is reduced by about 65% and is completely melted about 2 months earlier in the $2\times CO_2$ simulation. The reduced snow accumulation and earlier snowmelt decrease streamflow during the warm season under $2\times CO_2$. As will be discussed later, the change in the seasonal pattern of streamflow suggests a higher likelihood of wintertime flooding and reduced water supply in the summer under the greenhouse-warming scenario.

In the Middle Fork Flathead River (Fig. 2.3), the difference between the two scenarios is much less drastic as compared to the American River. Monthly maximum SWE is only reduced by 20% and there is little change in the timing of the peak streamflow. However, consistent with the warmer air temperature under $2\times CO_2$, streamflow during early winter is increased while that during late spring and summer is reduced, resulting in a slightly lower peak monthly flow.

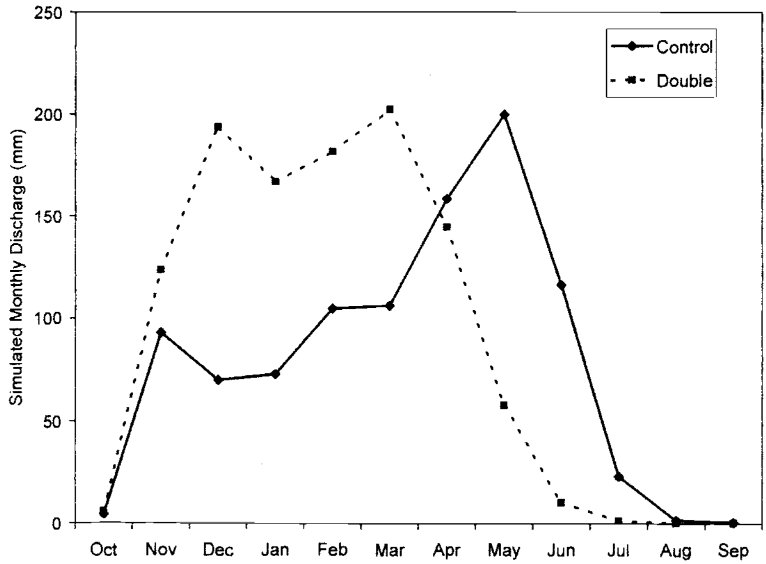


Fig. 2.2. Control and $2\times CO_2$ simulated discharge for the American River watershed.

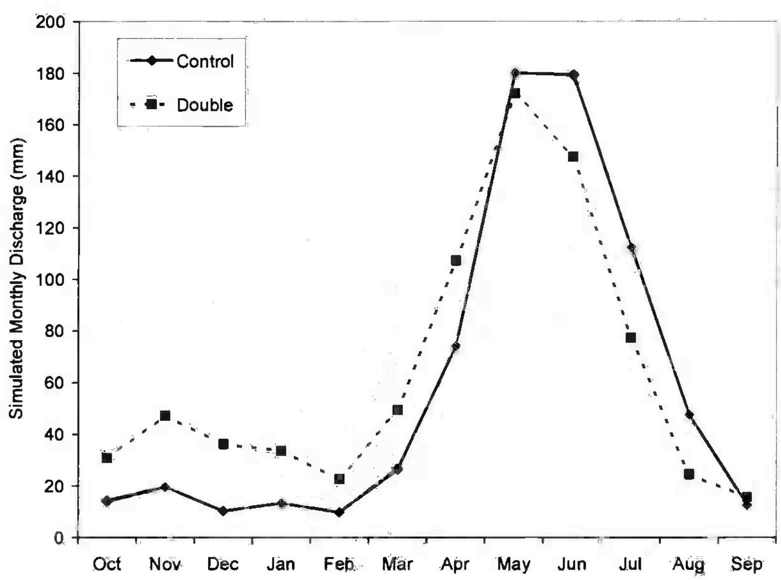


Fig. 2.3. Control and $2\times CO_2$ simulated discharge for the Middle Fork Flathead River watershed.

Flooding

Although the simulated increase in annual precipitation was less than 10% in both basins, there were significant changes in the frequency and intensity of flood events. Days with high precipitation typically occur from November to February in the American River basin, reflecting its maritime climate (Fig. 2.4). This pattern remains true under $2\times CO_2$ conditions, however the number of days with precipitation above the 1% daily control precipitation threshold is doubled (Table 2.2a). The mean intensity of these events is increased only 5% but the total volume is over twice that for the control climate, with the largest changes in November and December (Fig. 2.4).

Precipitation in the American River basin falls mostly as rain under $2\times CO_2$ climate, producing an increase in autumn and winter rain-on-snow events. The number of days with discharge above the 1% daily control discharge threshold increases about 70% under $2\times CO_2$ climate, while the total flow volume above this threshold increases by 80% (Table 2.3a and Fig. 2.5). The mean discharge of these floods increases about 9% from 21.9 mm d^{-1} to 23.9 mm d^{-1}. For comparison, the 5-year, 24-h return flow is 22.4 mm d^{-1}. In the control scenario, flooding generally occurs from November to February (as rain-on-snow) and/or during the snowmelt season in May and June (Fig. 2.5). The earlier loss of snowpack under the $2\times CO_2$ climate eliminates the May and June floods.

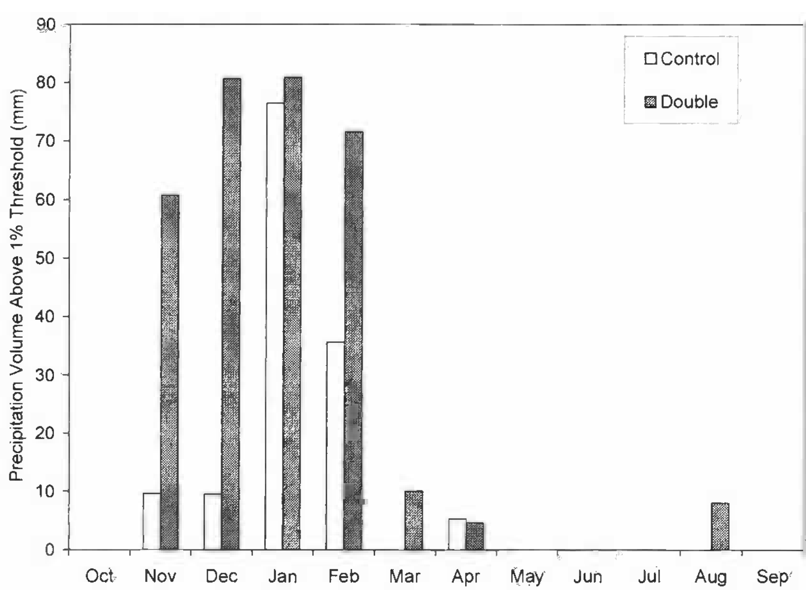


Fig. 2.4. Mean monthly volumes of precipitation above the 1% daily control precipitation threshold in the American River watershed under control and $2\times CO_2$ scenarios.

Table 2.2. Characteristics of precipitation above the 1% daily control precipitation threshold simulated for control and $2\times CO_2$ conditions in the (a) American River and (b) Middle Fork Flathead River watersheds.

Variable	Control simulation	$2\times CO_2$ simulation	Change
	(a) American River		
Days per year	3.6	7.9	4.3 (119%)
Mean intensity (mm d^{-1})	38.2	40.2	2.0 (5%)
Total volume (mm y^{-1})	137	317	180 (131%)
	(b) Middle Fork Flathead		
Days per year	3.6	4.0	0.4 (11%)
Mean intensity (mm d^{-1})	41.4	36.7	-4.7 (-11%)
Total volume (mm y^{-1})	148	147	-1 (-0.1%)

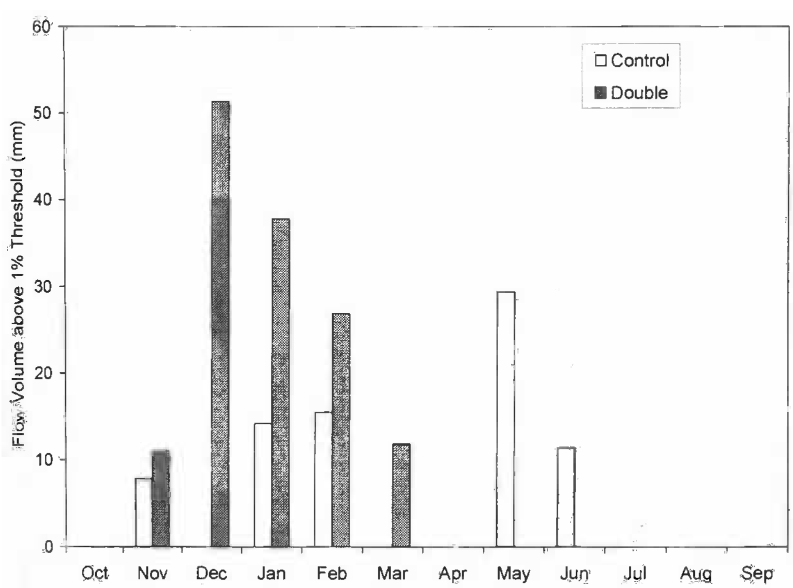


Fig. 2.5. Mean monthly flood volumes above the 1% daily control discharge threshold in the American River watershed under control and $2\times CO_2$ scenarios.

Table 2.3. Characteristics of flows above the 1% daily control discharge threshold simulated for control and $2\times CO_2$ conditions in the (a) American River and (b) Middle Fork Flathead River watersheds.

Variable	Control simulation	$2\times CO_2$ simulation	Change
	(a) American River[1]		
Days per year	3.6	6.0	2.4 (67%)
Mean discharge (mm d^{-1})	21.9	23.9	2.0 (9%)
Total volume (mm y^{-1})	78	139	61 (78%)
	(b) Middle Fork Flathead[2]		
Days per year	3.6	0.3	-3.3 (-92%)
Mean discharge (mm d^{-1})	15.8	13.0	-2.8 (-18%)
Total volume (mm y^{-1})	57	3.0	-54 (-95%)

[1]Based on 1% daily control discharge threshold of 17 mm d^{-1}. The 2- and 5-year, 24-h flows (based on 60 years of record) are 16.4 mm d^{-1} and 22.4 mm d^{-1}, respectively.

[2]Based on 1% daily control discharge threshold of 12.8 mm d^{-1}. The 1.25- and 2-year, 24-h flows (based on 58 years of record) are 11.9 mm d^{-1} and 15.5 mm d^{-1}, respectively.

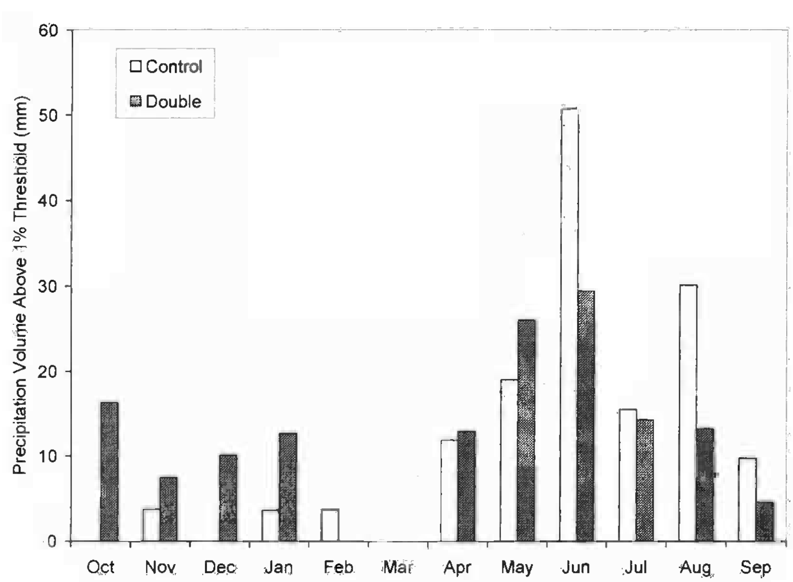


Fig. 2.6. Mean monthly volumes of precipitation above the 1% daily control precipitation threshold in the Middle Fork Flathead River watershed under control and $2\times CO_2$ scenarios.

The seasonal distribution of days with high precipitation for the control climate conditions in the Middle Fork Flathead River basin is reflective of its continental climate; most events occur in the spring and summer (Fig. 2.6). The number of days exceeding the 1% daily control precipitation threshold increases slightly in a 2×CO_2 climate with a small decrease in mean intensity and total volume (Table 2.2b). There is also a seasonal shift showing an increase in extreme precipitation during the autumn and winter (Fig. 2.6).

Flooding in the control climate scenario occurs during the snowmelt season with the highest volumes occurring from April to June (Fig. 2.7). Flooding is reduced significantly in a 2×CO_2 climate despite similar volumes of extreme precipitation (Table 2.3b). Flooding requires the right combination of a number of variables including antecedent moisture conditions, precipitation, net radiation, air temperature, and wind. Much of the reduction in flooding can be explained by differences in precipitation and air temperature. Average daily air temperature and precipitation during the top 1% of the control flows is 9.7°C and 18 mm d^{-1}, respectively. In contrast, the average daily air temperature and precipitation during the top 1% of flows in a 2×CO_2 climate is 8.5°C and 10 mm d^{-1}, respectively. Precipitation and air temperature did not occur in the right combinations in a 2×CO_2 climate simulation to produce the level of flooding observed in the control climate. In fact, the 1% daily control discharge threshold was exceeded only twice under 2×CO_2 climate.

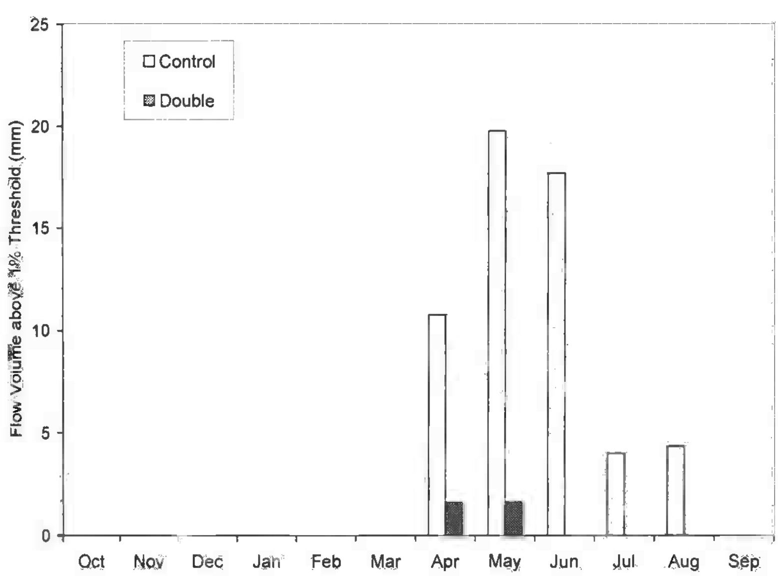


Fig. 2.7. Mean monthly flood volumes above the 1% daily control discharge threshold in the Middle Fork Flathead River watershed under control and 2×CO_2 scenarios.

Sensitivity analysis

A series of sensitivity scenarios were run to determine the relative significance of precipitation changes and air-temperature changes on the hydrologic changes associated with the alternative climates previously discussed. DHSVM was run for both basins using observed meteorology from Water Years 1989–1995. A series of runs were then made with a 15% decrease in daily observed precipitation, a 15% increase in daily observed precipitation, and increases to observed air temperature of 2°C and 4°C. The results are summarized in Tables 2.4 and 2.5 for the American and Middle Fork Flathead Rivers, respectively.

Table 2.4. Sensitivity analysis for the American River basin.

Precip. change	Temp change (°C)	ET (mm y^{-1})	Max SWE (mm y^{-1})	Q (mm y^{-1})	Days per year[1]	Mean discharge[1] (mm d^{-1})	Total volume[1] (mm y^{-1})
-15%	+ 0	780	547	739	2.0	19.2	38
-15%	+ 2	831	246	701	3.6	19.9	71
-15%	+ 4	887	105	654	4.1	20.7	86
+ 0%	+ 0	843	672	946	3.6	20.3	72
+ 0%	+ 2	899	307	903	5.7	22.2	127
+ 0%	+ 4	956	135	855	7.9	22.0	173
+15%	+ 0	899	799	1161	5.4	21.5	117
+15%	+ 2	959	376	1113	7.7	24.5	189
+15%	+ 4	1018	166	1063	12.3	23.4	287

[1]Above daily control discharge threshold of 14.0 mm d^{-1}. The 2- and 5-year, 24-h flows (based on 60 years of record) are 16.4 mm d^{-1} and 22.4 mm d^{-1}, respectively.

Table 2.5. Sensitivity analysis for the Middle Fork Flathead River basin.

Precip. change	Temp change (°C)	ET (mm y^{-1})	Max SWE (mm y^{-1})	Q (mm y^{-1})	Days per year[1]	Mean discharge[1] (mmn d^{-1})	Total volume[1] (mm y^{-1})
-15%	+ 0	508	459	604	0.7	11.1	8
-15%	+ 2	582	311	569	0.3	11.1	3
-15%	+ 4	655	173	535	0.0	0.0	0
+ 0%	+ 0	520	555	770	3.6	12.3	44
+ 0%	+ 2	599	389	729	0.9	13.1	11
+ 0%	+ 4	681	221	688	0.0	0.0	0
+15%	+ 0	528	652	941	9.7	12.8	124
+15%	+ 2	611	468	893	2.1	13.5	29
+15%	+ 4	696	269	846	0.4	18.5	8

[1]Above daily control discharge threshold of 10.7 mm d^{-1}. The 1.25-, 2-, and 5-year, 24-h flows (based on 58 years of record) are 11.9, 15.5, and 22.1 mm d^{-1}, respectively.

Simulation results for the American River with no changes in precipitation or air temperature are shown in Table 2.4, row 4. This simulation was used to establish the 1% daily discharge threshold of 14 mm d^{-1}. The influence of mean changes in precipitation without changes in mean air temperature can be seen by comparing rows one, four, and seven. As expected, both evapotranspiration and mean annual discharge increase with higher precipitation. There is also an increase in the frequency, intensity and total volume of flooding (columns 6–8). An increase in air temperature with precipitation held constant (rows 4–6) increases evapotranspiration and consequently reduces discharge. However, despite a decrease in mean annual flow, the frequency, intensity, and volume of flooding increases with a greater number of winter rain-on-snow events.

Simulation results for the Middle Fork Flathead River with no changes in precipitation or air temperature are shown in Table 2.5, row 4. This simulation was used to establish the 1% daily discharge threshold of 10.7 mm d^{-1}. The influence of mean changes in precipitation without changes in mean air temperature can be seen by comparing rows one, four, and seven. The trend is similar to the American River basin as both evapotranspiration and mean annual discharge increase with higher precipitation. Likewise, there is also an increase in the frequency, intensity, and total volume of flooding (columns 6–8). An increase in air temperature with precipitation held constant results in increased evapotranspiration and consequently a reduction in discharge. However, unlike the American River, the frequency and volume of flooding decreases with increasing air temperature. Winter rain-on-snow events produce the largest floods in the American River, whereas large floods in the Middle Fork Flathead basin are produced by spring rain falling on a ripe snowpack. Warming exacerbates winter rain-on-snow flooding in the American River, and reduces the magnitude of spring rain-on-snow events in both basins.

Conclusion

The effects of changes in climate on streamflow vary considerably between watersheds depending on the climatic regime and watershed characteristics. Our case study indicates that climate change will have important impacts on streamflow and flooding in forested mountain watersheds in the Pacific Northwest of the United States. There will be a higher likelihood of winter rain-on-snow flooding in coastal mountain watersheds such as the American River. This may be accompanied by an increased incidence of landslides and debris flows along with changes in stream channel morphology. Reduced soil moisture in the summer may lead to increased fire hazard and reduced forest growth.

The difference between the impacts in maritime versus continental watersheds is very large, thus highlighting the regional nature of the problem. In the Northern Rockies, where winter air temperature is sufficiently low even under the $2\times CO_2$ conditions, the impacts of climate change on streamflow are modest. The seasonal pattern of streamflow remains intact and the incidence of flooding is reduced. In maritime watersheds, however, the cold-season air temperature over much of the area is just around freezing. A shift in the freezing level combined with earlier snowmelt caused by the warming can trigger a significant shift in the seasonal pattern of streamflow and the frequency of flooding.

Although climate model predictions of precipitation changes under greenhouse warming are highly uncertain, this study suggests that the response in snowpack and streamflow over the Pacific Northwest is largely determined by air temperature changes. However, one should be cautioned that this conclusion is derived based on results from snow-dominated watersheds. For rain-dominated or mixed rain/snow type watersheds the climate impacts are likely to be highly dependent on both air temperature and precipitation changes.

References

Gleick, P.H. (1999) Introduction: Studies from the water sector of the National Assessment. *Journal of American Water Resources Association* 35, 1297-1300.

IPCC (Intergovernmental Panel on Climate Change) (1996) *The Science of Climate Change*. Houghton, J.T., Meira Filho, L.G., Callander, B.A., Harris, N., Kattenberg, A. and Maskell, K. (eds). Cambridge University Press, New York.

Kiehl, J.T., Hack, J.J., Bonan, G.B., Boville, B.A., Briegleb, B.P., Williamson, D.L. and Rasch, P.J. (1996) Description of the NCAR Community Climate Model (CCM3), *NCAR Tech. Note, NCAR/TN-288+IA*. National Center For Atmospheric Research, Boulder, Colorado.

Lettenmaier, D.P. and Gan, T.Y. (1990) Hydrologic sensitivities of the Sacramento-San Joaquin River basin, California, to global warming. *Water Resources Research* 26, 69-86.

Leung, L.R. and Ghan, S.J. (1995) A subgrid parameterization of orographic precipitation. *Theoretical and Applied Climatology* 52, 95-118.

Leung, L.R. and Ghan, S.J. (1998) Parameterizing subgrid orographic precipitation and land surface cover in climate models. *Monthly Weather Review* 126, 3271-3291.

Leung, L.R. and Ghan, S.J. (1999a) Pacific Northwest climate sensitivity simulated by a regional climate model driven by a GCM. Part I: Control simulations. *Journal of Climate* 12, 2010-2030.

Leung, L.R. and Ghan, S.J. (1999b) Pacific Northwest climate sensitivity simulated by a regional climate model driven by a GCM. Part II: $2\times CO_2$ simulations. *Journal of Climate* 12, 2031-2058.

Leung, L.R. and Wigmosta, M.S. (1999) Potential climate change impacts on mountain watersheds in the Pacific Northwest. *Journal of American Water Resources Association* 35, 1463-1471.

Leung, L.R., Wigmosta, M.S., Ghan, S.J., Epstein, D.J. and Vail, L.W. (1996) Application of a subgrid orographic precipitation/surface hydrology scheme to a mountain watershed. *Journal of Geophysical Research* 101, 12803-12817.

Manabe, S., Stouffer, R.J., Spelman, M.J. and Bryan, K. (1991) Transient responses of a coupled ocean-atmosphere model to gradual changes of atmospheric CO_2. Part I: Annual mean response. *Journal of Climate* 4, 785-818.

Storck, P., Bowling, L., Weatherbee, P. and Lettenmaier, D.P. (1998) Application of a GIS-based distribution hydrology model for prediction of forest harvest effects on peak stream flow in the Pacific Northwest. *Hydrological Processes* 12, 889-904.

Wigmosta, M.S. and Lettenmaier, D.P. (1994) A distributed hydrologic model for mountainous catchments. In: Rosso, R., Peano, A., Becchi, I. and Bemporad, G.A. (eds), *Advances in Distributed Hydrology*. Water Resources Publications, Fort Collins, Colorado, pp. 359-378.

Wigmosta, M.S. and Lettenmaier, D.P. (1999) A comparison of simplified methods for routing topographically-driven subsurface flow. *Water Resources Research* 35, 255-264.

Wigmosta, M.S., Vail, L.W. and Lettenmaier, D.P. (1994) A distributed hydrology-vegetation model for complex terrain. *Water Resources Research* 30, 1665-1679.

Chapter 3

Regional Hydrologic Impacts of Climatic Change

Donald H. Burn
Department of Civil Engineering, University of Waterloo, Waterloo, Ontario, Canada

A climate change detection framework is presented and applied to five hydrologic variables calculated for a collection of forested catchments from three climatic regions in Canada. The trend detection framework includes the application of a non-parametric statistical test for trend and the use of an exploratory data analysis technique. The application of the trend detection technique to time series of hydrologic variables for the forested catchments reveals that a greater number of trends are identified than are expected to occur by chance. Mapping of the trend results indicates clear spatial patterns in the nature and the direction of the trends that were identified as well as differences in results between the climate regions investigated. The Yukon/North British Columbia Mountains (YNBC) climatic region exhibited increasing flow trends in two low indices, the annual flow, and to a lesser extent, in peak flows. This result is consistent with climate change predictions for northern latitude regions. In contrast, decreasing flow conditions are expected in the other two Canadian forest climatic regions: North-East Forest and North-West Forest.

Introduction

The potential impacts of climatic change have recently received a great deal of attention from researchers in a variety of fields. Reviews of related work include Gleick (1989) and the Intergovernmental Panel on Climate Change (IPCC) (1996). From these and other works, a range of potential impacts on the hydrologic regime for various geographic areas has been hypothesized. Areas that could be negatively impacted in terms of water supply availability have been identified (Revelle and Waggoner, 1983; Gleick, 1987) as well as locations likely to experience changes in runoff production (Gleick, 1986; Lettenmaier and Gan, 1990).

A comprehensive review of the potential impacts of climatic change is provided in IPCC (1996). This report indicates that climatic change is likely to increase runoff in higher latitude regions because of increased precipitation. Changes in flood frequencies are expected in some locations, particularly in northern latitudes and in catchments experiencing snowmelt flooding events. The frequency and severity of drought events could increase as a result of changes in both precipitation and evapotranspiration. Changes in the hydrologic regime that do occur are not expected to be equally distributed throughout the year. For example, increased temperatures in the winter are expected to lead to earlier snowmelt events and a shift in runoff from the spring to late winter with a corresponding decrease in runoff in the summer period. This in turn could lead to water shortages for areas currently experiencing demands that are large in comparison to the available water resources (for example, parts of the Canadian prairies).

Burn (1994a) identified a trend in the timing of the peak snowmelt runoff event for catchments in west-central Canada with more recent events occurring earlier in the year. Lins and Michaels (1994) identified increases in the autumn and winter streamflow in the United States. These increases were related to changes in the temperature. Increases in winter and spring streamflow for much of the United States were found by Lettenmaier *et al.* (1994). Westmacott and Burn (1997) found decreases in streamflow for the Canadian prairies that could be related to changes in the temperature.

The focus in the work described herein is on climate change impacts for forested catchments. IPCC (1996) indicates that forests may be particularly vulnerable to extremes in the availability of water. The expected impacts of climate change on forested catchments are projected to differ for different types of forests. Tropical forests are expected to be most affected by changes in the soil water availability. Temperate forests are projected to experience significant changes in species composition as a result of changing climatic conditions. Temperate forests may be lost, in regions that are already short of water, because of drought conditions in the summer. Boreal forests are expected to be most strongly affected by temperature changes because of their high latitude locations.

It is difficult to predict the impacts of climate change on the hydrology of forested catchments because of the complex processes that are involved in runoff generation in a forested catchment. Running and Nemani (1991) report on the modelling of the water balance for a forested area in Montana under climate change conditions. Running and Nemani (1991) found that runoff from the forested area would decrease dramatically primarily because of a decrease in the snowmelt contribution. The same study found that substantially different responses could be expected from forested catchments from different geographic areas.

This chapter identifies some of the hydrologic impacts of climatic change for forested catchments in Canada. The emphasis in the research reported herein is on the quantification of trends in hydrologic variables and the investigation of the relationship between trends in hydrologic variables and trends in temperature. The spatial distribution of catchments exhibiting trends and not exhibiting trends is also investigated. The next section of this chapter describes the hydrologic variables that are examined in this work. This is followed by a presentation of the trend detection techniques that are used. The methodology outlined is then applied to a collection of forested catchments in Canada and the chapter ends with a summary and conclusions for selected climatic regions in Canada.

Hydrologic Variables

There have been numerous studies conducted investigating the impacts of climatic change on water resources. From these studies, many different variables have been suggested as potential indicators of climate change. Schadler (1987) considered the use of components of the water balance for monitoring climate change. The use of proxy data, such as tree ring growth rates or information from ice cores, was suggested by Lawford (1988) as an approach to monitor climate change. The use of streamflow as the hydrologic variable of interest was advocated by Pilon *et al.* (1991). Anderson *et al.* (1992) considered measures of the low, high and average flow regime for climate change detection. Ideally, numerous variables should be monitored due to the anticipated diversity of responses to climatic change. Some of the issues to consider when selecting variables include the natural variability of the monitoring variables, the spatial coverage obtained with the variables selected, and the record length available for establishing baseline conditions and any subsequent changes or deviations.

Hydrologic variables are selected for identifying the impacts of climatic change in the work reported herein due to the reasonable data availability, both temporally and spatially, and the fact that these variables are of direct interest in water resources management. Climatic change is expected to affect the magnitude, frequency, and timing of hydrologic events. It is therefore important that the selection of response parameters reflects various components of the hydrologic regime. Components selected should ideally include measures of (Burn, 1994b): (i) the average flow regime; (ii) the annual and seasonal extremes in the flow regime; (iii) the seasonal distribution of the runoff; (iv) the variability of the runoff response; and (v) the timing of hydrologic events. Including a diversity of response measures should lead to a more comprehensive climate-change detection approach and should result in a better prediction of the range of climatic change impacts that could occur.

Five hydrologic variables were selected for analysis in this work. The variables selected include: (i) a base flow index, BFI, which expresses the ratio of the volume of the base flow to the annual flow volume, as a fraction; (ii) the annual 7-day low flow, Q7; (iii) the annual peak daily flow, Q_p; (iv) the mean annual flow, Q_a; and (v) the coefficient of variation of the daily flow values for the year, CV. The base flow was calculated using a hydrograph separation approach that is based on a filtering technique. The first two variables (BFI and Q7) reflect different aspects of the low flow regime, the third (Q_p) is a measure of the high flow (flood) regime, the fourth (Q_a) is a measure of the overall availability of water and the fifth variable (CV) is a measure of the variability of the hydrologic regime.

Impact Detection Technique

When attempting to detect climatic signals in a natural series one must be cognizant of the inherent variability of hydrologic time series (Burn, 1994c). Askew (1987) indicates that there is a difficulty associated with differentiating between natural variability and the impacts of climatic change. The difficulty occurs because the signal that results from climate change must be separated from the noise that is a natural part of the hydrologic record. This argues for the development of a rigorous procedure for detecting climatic change impacts.

The impact detection procedure used herein involves a combination of data analysis approaches. A statistical test for trend is used in order to identify trends in time series of hydrologic event magnitudes. In conjunction with the trend test, exploratory data analysis techniques are used to examine the relationship between the hydrologic variables and climatic change. Exploratory data analysis techniques are designed to help uncover any underlying relationships that exist amongst a set of variables. In this work, a smoothing technique, referred to as LOWESS (Cleveland, 1979), is used to examine the behaviour of variables over time and to identify relationships between pairs of variables. In particular, the smoothing technique is used to explore the pattern of hydrologic variables versus time. The resulting time series plots are used to determine the overall tendency of the series describing one of the hydrologic variables at a given location. The technique is also used to determine if there is a relationship between patterns in hydrologic variables and patterns in a climate (temperature) series. The intent with exploratory data analysis is to aid in the understanding of the behaviour of one or more variables through a visual display of important characteristics of the data set.

The statistical trend test selected was the Mann-Kendall non-parametric test for trend (Mann, 1945; Kendall, 1975). This trend test has been used by other researchers in similar applications (Hirsch *et al.*, 1982; Gan and Kwong, 1992) and has been found to be an effective tool for identifying trends in hydrologic and other related variables. The test statistic for the Mann-Kendall test is given as

$$S = \sum_{k=1}^{n-1} \sum_{j=k+1}^{n} \mathrm{sgn}(x_j - x_k)$$

where the x_j are the sequential data values, n is the length of the data set, and sgn (θ) is the sign function defined as

$$\operatorname{sgn}(\theta) = \begin{cases} 1 & \text{if } \theta > 0 \\ 0 & \text{if } \theta = 0 \\ -1 & \text{if } \theta < 0 \end{cases}$$

where θ is the argument of the sign function.

The theoretical mean and variance of the test statistic, under the null hypothesis of no trend in the series, are given as

$$E[S] = 0$$

and

$$\operatorname{Var}[S] = \frac{n(n-1)(2n+5) - \sum t(t-1)(2t+5)}{18}$$

where t is the extent of any tie (i.e., the number of data points involved in a tie) and the summation is over all ties. For sample sizes larger than ten, the statistic is very nearly normally distributed if a continuity correction is applied giving

$$S' = S - \operatorname{sgn}(S)$$

where, S' is the corrected test statistic value. A Z value associated with the trend statistic can be calculated, assuming the corrected test statistic follows the normal distribution, as follows

$$Z = \frac{S'}{\sqrt{Var[S]}}$$

where Z is a standard normal variate. The magnitude of the Z value obtained can then be used to determine the significance of any trend in the data set. It is also possible to obtain a non-parametric estimate for the magnitude of the slope following Hirsch *et al.* (1982):

$$\beta = \text{Median}\left\{\frac{x_j - x_k}{j-k}\right\} \qquad \text{for all} \quad k < j$$

where β is a robust estimate of the slope.

The procedure adopted to detect the impacts of climatic change is as follows. The Mann-Kendall test is first applied to each data set. There are five data sets for each catchment analysed with one data set corresponding to each of the hydrologic variables. The LOWESS smoothing technique is applied to each time series and is also applied to an annual temperature series for the climatic region for the catchment. Finally, the correlation between the smoothed time series for the hydrologic variable and the smoothed time series for the regional temperature series is evaluated and catchments exhibiting a significant correlation are noted. The correlation is evaluated using Kendall's tau, which is a non-parametric estimate of the correlation. Kendall's tau is calculated (Snedecor and Cochran, 1980) by assigning a rank to each x and y observation, where the x's and y's represent data pairs. The data pairs are then ordered in accordance with the ranking of one of the two variables (for example, the x variable). The rankings of the other variable in the data pair (for example, the y variable) are then used in the following way. For each observation, the number of observations that follow the current observation and have a smaller rank is determined. The sum of these numbers is called the count, Q. Kendall's tau is then defined as

$$\tau = 1 - \frac{4Q}{n(n-1)}$$

where n is the number of data pairs. Kendall's tau gives similar results to other non-parametric estimates of the correlation, such as Spearman's rank correlation coefficient, but its frequency distribution for the null hypothesis of no correlation is simpler.

In the previously outlined procedure, the correlation is calculated between the smoothed hydrologic variable and the smoothed temperature series. The smoothing is applied prior to calculating the correlations since the intent is to identify any similarities in the underlying pattern in each variable. This type of relationship can be difficult to detect in data that have not been smoothed as a result of, for example, a lag in the variable response between the data sets (Westmacott and Burn, 1997).

Application to Canadian Catchments

Description of study area

The catchments selected for analysis in this work had to meet a number of criteria. First, they had to be from a forested area given the focus of this work on climate change impacts for forested catchments. Second, the catchments had to have no significant regulation (e.g., dams) or diversions. Third, the catchments had to represent pristine conditions, or at least reflect relatively stable land use during the gauging period. Fourth, the record length available for the catchment had to be sufficiently long to enable statistical tests to be reliably performed on the data record. Fifth, the hydrometric data collected for the catchment had to be considered accurate. Sixth, the gauging station for the catchment had to be an active gauging station that is currently collecting streamflow data. All of the catchments selected for analysis were drawn from the Reference Hydrometric Basin Network (RHBN), which is a collection of catchments recently identified by Environment Canada (1999) for use in climate change studies. Catchments in the RHBN satisfy the second, third, fifth and sixth criteria noted above. The minimum record length for the catchments in the RHBN is 20 years. This was felt to be too short for the purposes of this work so only those catchments with a minimum record length of 30 years were selected. A Geographic Information System (GIS) was used to identify those catchments from the RHBN that could be considered to represent forested catchments. All of the catchments that were selected are part of the boreal forest region of Canada.

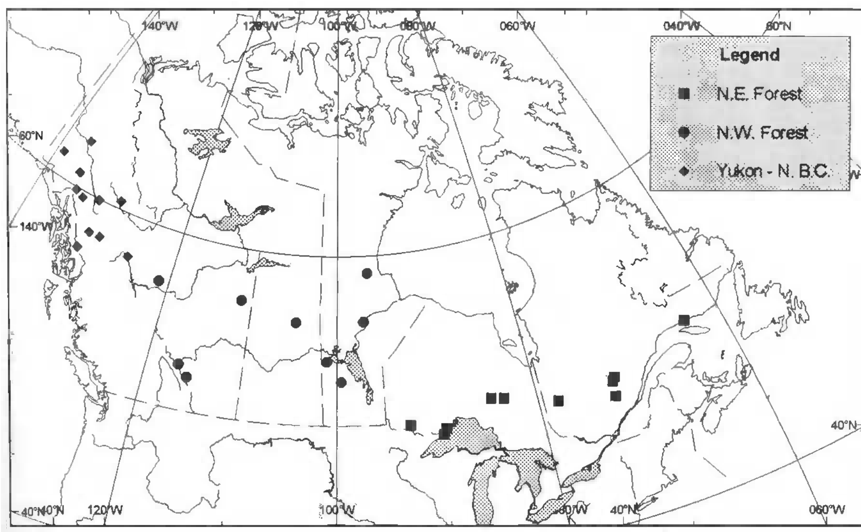


Fig. 3.1. Location of the catchments and the three forested climatic regions investigated.

The application of the above criteria to the available database resulted in the identification of 30 catchments with the desired characteristics. The locations of the catchments are shown in Fig. 3.1. Fig. 3.1 additionally identifies the climatic region for each of the catchments. The climatic regions are based on a classification by Gullet (1992) and reflect the different climatic conditions within different geographic areas of Canada. The catchments that were selected for analysis are drawn from three of the 11 climatic regions that have been identified for Canada.

Table 3.1. Characteristics of the catchments included in the analysis.

Hydrometric station number	River name	Province or territory[1]	Climate region[2]	Drainage area (km^2)	Years of record
02AA001	Pigeon River	ON	NEF	1,550	50
02AB008	Neebing River	ON	NEF	187	42
04JC002	Nagagami River	ON	NEF	2,410	45
04LJ001	Missinaibi River	ON	NEF	8,940	50
05PB014	Turtle River	ON	NEF	4,870	47
02NE011	Croche Riviere	QC	NEF	1,570	30
02RD002	Mistassibi Riviere	QC	NEF	9,320	35
02RF001	Chamouchouane Riviere	QC	NEF	15,300	33
02VC001	Romaine Riviere	QC	NEF	13,000	38
04NA001	Harricana Riviere	QC	NEF	3,680	50
05BB001	Bow River	AB	NWF	2,210	50
05DA007	Mistaya River	AB	NWF	249	30
07CD001	Clearwater River	AB	NWF	30,800	38
05LD001	Overflowing River	MB	NWF	3,350	35
05LH005	Waterhen River	MB	NWF	55,000	43
05TD001	Grass River	MB	NWF	15,400	35
06GD001	Seal River	MB	NWF	48,100	37
06CD002	Churchill River	SK	NWF	119,000	32
07FC003	Blueberry River	BC	NWF	1,750	30
07EA002	Kwadacha River	BC	YNBC	2,410	34
08CC001	Klappan River	BC	YNBC	3,550	31
08CE001	Stikine River	BC	YNBC	29,300	31
08CG001	Iskut River	BC	YNBC	9,350	31
09AA006	Atlin River	BC	YNBC	6,810	45
09AA010	Lindeman Creek	BC	YNBC	240	35
09AE003	Swift River	BC	YNBC	3,320	37
09AC001	Takhini River	YT	YNBC	6,990	44
09BC001	Pelly River	YT	YNBC	49,000	41
09CA002	Kluane River	YT	YNBC	4,950	41
10AB001	Frances River	YT	YNBC	12,800	31

[1]ON indicates Ontario, QC denotes Quebec, AB denotes Alberta, MB indicates Manitoba, SK indicates Saskatchewan, BC denotes British Columbia and YT denotes Yukon Territory.
[2]NEF denotes North-East Forest, NWF indicates North-West Forest, and YNBC denotes Yukon/North BC Mountains.

Table 3.1 summarizes the characteristics of the catchments. Shown in this table is the province in which the catchment is located, the climatic region for each catchment, the drainage area for the catchment, and the years of record used in the analysis. As noted previously, the selection criteria for choosing catchments to analyse included a lower bound on the record length of 30 years.

In addition to this lower bound, 50 years was selected as an upper bound for analysis such that for any catchment with a record length of greater than 50 years, only the most recent 50 years were analysed. The upper bound of 50 years was adopted since it was felt that a longer record length could mask trends that have arisen only in the most recent portion of the data record. The upper bound of 50 years also provides less diversity in the periods of record analysed for the collection of 30 catchments. The 30 catchments have a range of drainage areas from 187 to 119,000 km^2 with a median value of 5880 km^2. The median record length is 37 years with a range of 30 to 50 years.

Presentation of results

The results from the application of the trend test to the hydrologic variables are summarized in Table 3.2. Shown in Table 3.2 is the number (and percentage) of catchments exhibiting a trend for each of the hydrologic variables considered. Results are presented separately for the entire data set and for each of the three climatic regions. Three significance levels have been selected. A significance level of greater than 10% indicates that the series does not exhibit a trend, a significance level of 10% indicates at least moderate evidence of a trend, and a significance level of 5% indicates strong evidence of a trend in the series. Note that those catchments with a significant trend at the 5% level are a subset of the catchments with a significant trend at the 10% level.

Table 3.2. Number (and percentage) of catchments in each trend test significance level class.

Data set	Significance level	Hydrologic variable				
		Base flow index (BFI)	7-Day low flow (Q7)	Peak flow (Q_p)	Annual flow (Q_a)	Coefficient of variation (CV)
All data	> 10%	19 (63%)	18 (60%)	20 (67%)	20 (67%)	19 (63%)
	10%	11 (37%)	12 (40%)	10 (33%)	10 (33%)	11 (37%)
	5%	7 (23%)	11 (37%)	8 (27%)	5 (17%)	6 (20%)
NEF	> 10%	6 (60%)	8 (80%)	6 (60%)	8 (80%)	4 (40%)
	10%	4 (40%)	2 (20%)	4 (40%)	2 (20%)	6 (60%)
	5%	2 (20%)	2 (20%)	3 (30%)	0 (0%)	4 (40%)
NWF	> 10%	6 (67%)	6 (67%)	5 (56%)	6 (67%)	9 (100%)
	10%	3 (33%)	3 (33%)	4 (44%)	3 (33%)	0 (0%)
	5%	2 (22%)	3 (33%)	4 (44%)	3 (33%)	0 (0%)
YNBC	> 10%	7 (64%)	4 (36%)	9 (82%)	6 (55%)	6 (55%)
	10%	4 (36%)	7 (64%)	2 (18%)	5 (45%)	5 (45%)
	5%	3 (27%)	6 (55%)	1 (9%)	2 (18%)	2 (18%)

The trend test results for the baseflow index (BFI) are consistent across the three climatic regions in terms of the percentage of catchments that exhibit either a moderate or a strong trend. The geographic distribution of the catchments that exhibit trends is also fairly uniform as can be seen from Fig. 3.2. Of the 11 catchments exhibiting a trend, all but one of the catchments exhibit an increasing trend.

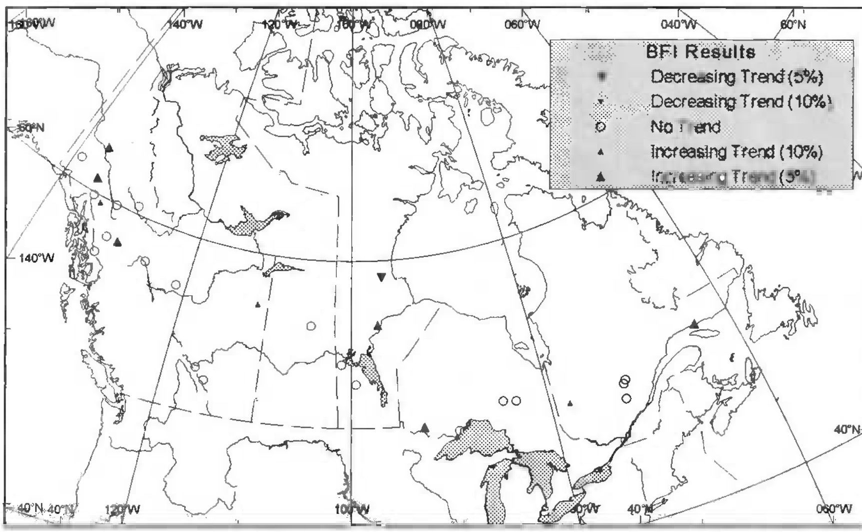


Fig. 3.2. Trend test results for BFI.

The 7-day low flow (Q7) variable results in both the largest number of trends and the largest number of trends significant at the 5% level in comparison with the other hydrologic variables. The Yukon/North BC Mountains (YNBC) climatic region exhibits the greatest percentage of significant trends while the North-East Forest (NEF) climatic region exhibits the fewest. Fig. 3.3 shows the geographic distribution of the catchments exhibiting a trend. As can be seen in Fig. 3.3, the trends all increase in the YNBC climatic region while only decreasing trends occur in the other two climatic regions.

The annual peak daily flow (Q_p) variable results in comparatively few trends in the YNBC climatic region with roughly an equivalent number in the remaining two climatic regions. Fig. 3.4 reveals that all of the trends in the YNBC climatic region are increasing trends while all of the other trends, with the exception of one series in the NEF climatic region, are decreasing trends.

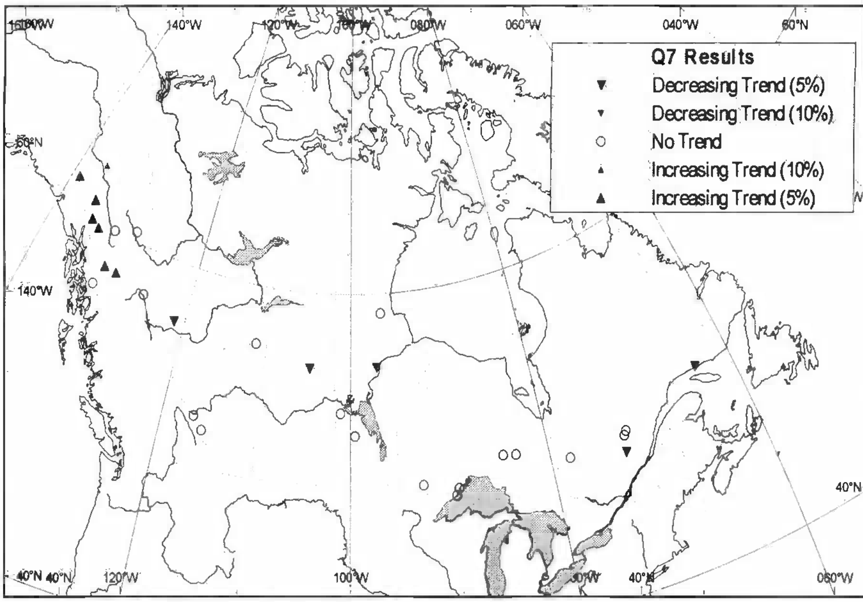


Fig. 3.3. Trend test results for Q7.

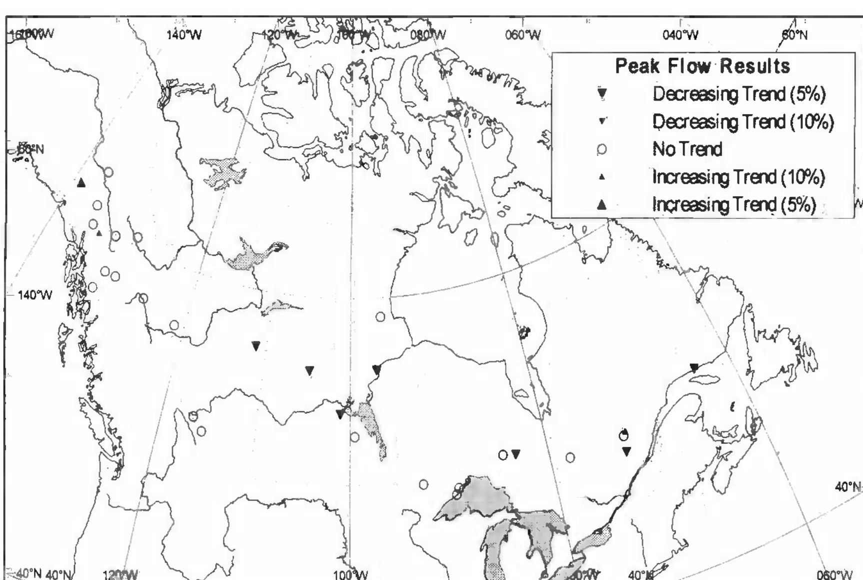


Fig. 3.4. Trend test results for peak flow, Q_p.

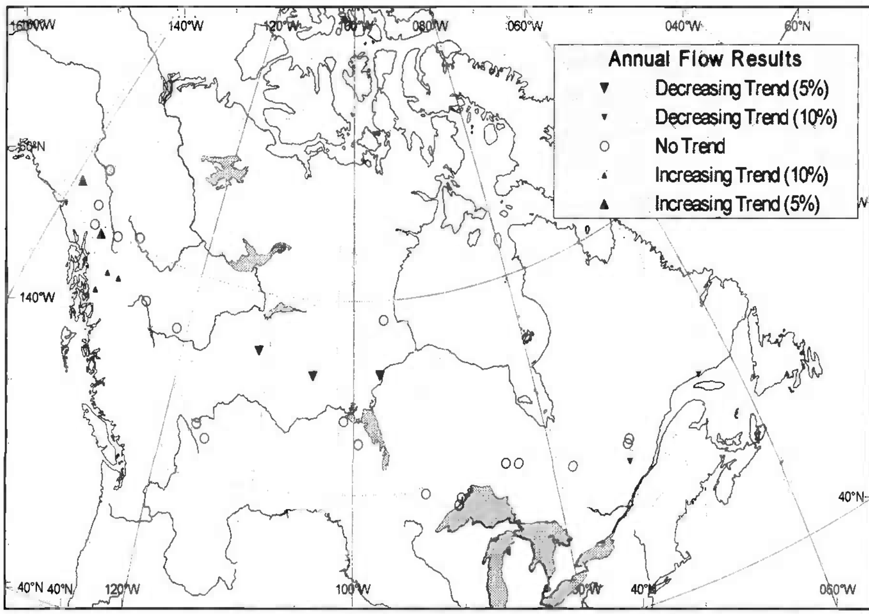


Fig. 3.5. Trend test results for annual flow, Q_a.

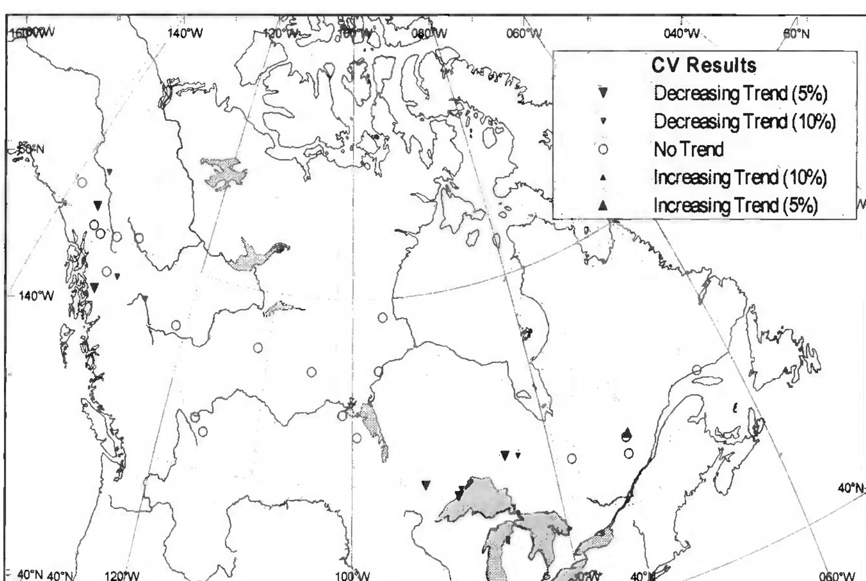


Fig. 3.6. Trend test results for CV.

Results for the annual flow (Q_a) variable show that the largest percentage of trends occurs in the YNBC climatic region with very few trends in the NEF climatic region. Of the two trends in the NEF climatic region, neither is significant at the 5% level. Fig. 3.5. reveals that the catchments in the YNBC climatic region exhibit increasing trends for the annual flow while all other catchments exhibit decreasing trends. The annual flow variable exhibits the fewest trends that are significant at the 5% level.

The results for the coefficient of variation of the daily flow series (CV) are noteworthy in that there are no significant trends for the catchments in the North-West Forest (NWF) climatic region. The NEF climatic region has the largest percentage of significant trends. Fig. 3.6 reveals that all but one of the trends from all three regions are decreasing trends.

The trend test was also applied to a regional annual temperature series for each climatic region. The regional temperature series represents the annual departure of the regional average temperature from a 30-year climate normal. The length of the data set analysed is 50 years, corresponding to the longest time period used for analyzing the hydrologic variables. The results are displayed in Table 3.3 that shows, for each climatic region, the significance level of the trend and the value for the non-parametric estimate of the slope. A highly significant trend exists in the temperature series from the NWF and YNBC climatic regions, but no evidence of a trend is apparent in the NEF climatic region. Both of the significant regional temperature trends are increasing.

Table 3.3. Trend test results for the annual temperature series for each climatic region.

Climatic region	Significance level	Slope estimate (°C year $^{-1}$)
NEF	> 10%	-0.006
NWF	1%	0.025
YNBC	0.5%	0.039

Table 3.4 displays results from the calculation of the correlation between each of the hydrologic variables and the corresponding regional temperature series. Table 3.4 presents the number (and percentage) of catchments that demonstrate a correlation that is significant at the 5% level. The correlations are estimated using Kendall's tau, a non-parametric estimate of the correlation. The correlations are calculated between the smoothed series, derived using LOWESS, for the hydrologic variable and the smoothed regional temperature series. The results reveal that the majority of the catchments demonstrate a significant correlation implying that the trends identified in Table 3.2 are associated with changes in the temperature. In comparing the hydrologic variables, it is apparent that the correlation is weakest for the annual flow variable and fairly consistent for the other variables. The correlations tend to be weakest for the NEF catchments and strongest for the YNBC catchments. Fig. 3.7 shows the number of variables for which there is a significant correlation for each of the catchments. The most notable feature of this map is the concentration

in the YNBC climatic region of catchments with a large number of variables that exhibit a significant correlation with the temperature series.

Table 3.4. Number (and percentage) of catchments with correlation between the hydrologic variable and the temperature series that is significant at the 5% level.

Data set	Hydrologic variable				
	Base flow index (BFI)	7-Day low flow (Q7)	Peak flow (Q_p)	Annual flow (Q_a)	Coefficient of variation (CV)
All data	22 (73%)	22 (73%)	20 (67%)	18 (60%)	21 (70%)
NEF	7 (70%)	5 (50%)	5 (50%)	4 (40%)	6 (60%)
NWF	8 (89%)	7 (78%)	6 (67%)	7 (78%)	4 (44%)
YNBC	7 (64%)	10 (91%)	9 (82%)	7 (64%)	11 (100%)

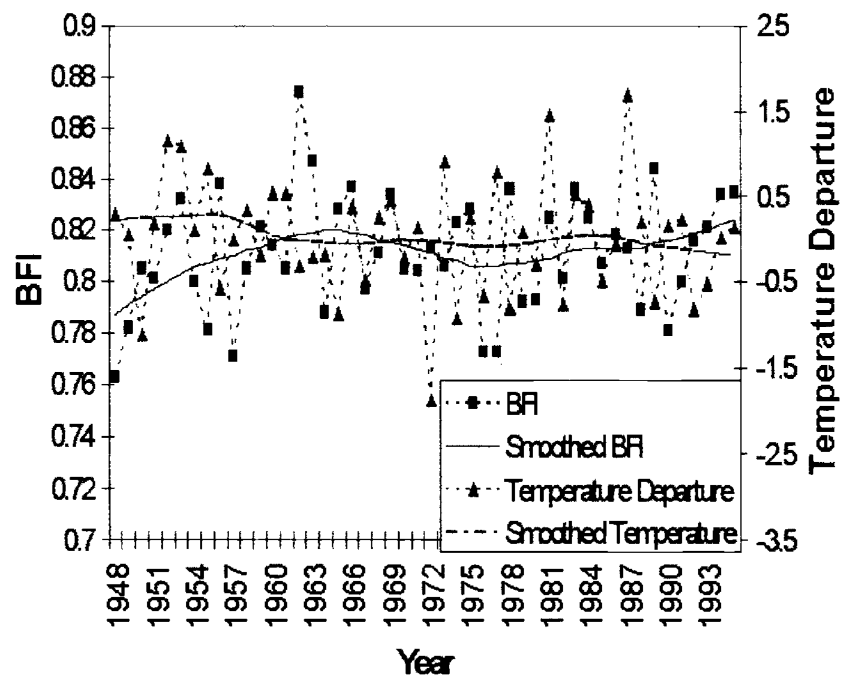


Fig. 3.8. Plot of BFI and the regional temperature series for a catchment in the NEF climatic region.

The final results to be presented are selected plots of time series for both a hydrologic variable and the corresponding regional temperature series. Figs. 3.8-3.10 show one example plot for a catchment selected from each climatic region.

In each case, the smoothed representation of each series is shown in addition to the observed values for each series. Fig. 3.8 shows the results for the BFI for the Pigeon River (02AA001), a catchment from the NEF climatic region. This catchment has an increasing trend in the BFI variable that is significant at the 10% level. The correlation with the regional temperature series is -0.532, which has a significance level of less than 5%. The inverse relationship between the BFI and temperature series implied by the negative correlation is apparent in the smoothed plots in Fig. 3.8. The inverse relationship between BFI and temperature implies that increases in temperature are associated with decreases in the BFI.

Fig. 3.9 shows the plots for the peak flow, Q_p, for the Grass River (05TD001), a catchment from the NWF climatic region. This catchment has a decreasing trend in Q_p that is significant at the 5% level and has a correlation with the regional temperature series of -0.779 (significance level is less than 5%). The very strong correlation between the smoothed temperature series and the peak flow series is apparent in the near mirror image displayed in Fig. 3.9. This is most apparent in the part of the series prior to around 1985 after which the peak flow series appears to decrease at a greater rate than would be expected based on the relationship with the temperature series in the earlier part of the series.

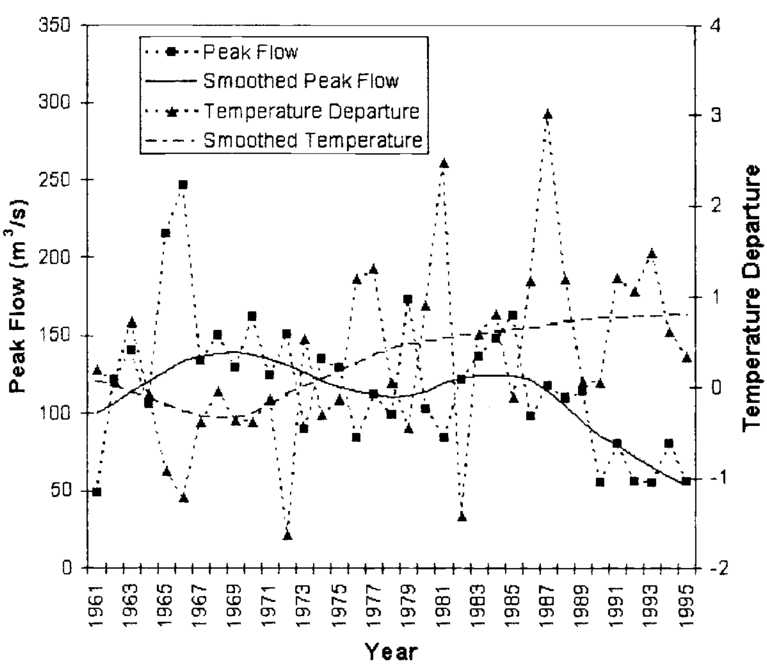


Fig. 3.9. Plot of peak flow and the regional temperature series for a catchment in the NWF climatic region.

Fig. 3.10 displays the plots for the 7-day low flow, Q7, for the Atlin River (09AA006), a catchment in the YNBC climatic region. This series has an increasing trend in Q7 that is significant at the 5% level. The correlation between the smoothed Q7 series and the smoothed regional temperature series is 0.781, which has a significance level of less than 5%. The strong relationship between the smoothed Q7 series and the smoothed temperature series is apparent in the plot in Fig. 3.10. With the exception of a brief period during the early 1960s, the correspondence in the patterns for the two smoothed series is very strong. The increasing trend in Q7 could be a result of changes in the hydrologic regime resulting in a shift in the timing of low flow events. The observed trend may also be partially attributable to changes in the precipitation regime.

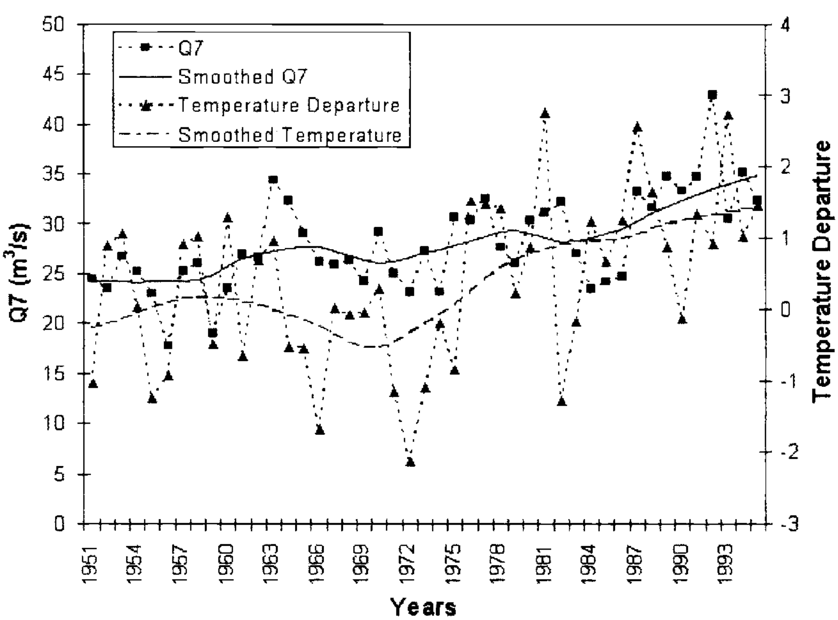


Fig. 3.10. Plot of Q7 and the regional temperature series for a catchment in the YNBC climatic region.

Discussion of results

The number of trends identified in the hydrologic variables for the 30 catchments is in excess of what would be expected to occur by chance. All of the values for the number of trends that are reported in Table 3.2 exceed the nominal number that are expected to occur by chance with the exception of Q_a for the NEF and CV for the NWF. Lettenmaier *et al.* (1994) note that the effect of cross-correlation in the data is to increase the expected number of trends under the hypothesis of no trend in the data. Livezey and Chen (1983) indicate the need to consider the field significance in ascertaining the overall significance of the outcomes from a set of statistical tests. Field significance allows the determination of the fraction of tests which are expected to show a trend, at a given local (nominal) significance level, purely by chance. The procedure involves first determining the significance of the results assuming spatial independence and then determining the effective sample size (accounting for cross-correlation) required to obtain overall significance. Applying this approach to the results for the entire data set of 30 catchments indicates that field significance is established, assuming spatial independence. The required number of significant outcomes at the 5% level is five and at the 10% level the required number is seven. These values equal or exceed the minimum observed values from Table 3.2 of five and ten, respectively. Accounting for spatial dependence gives effective sample sizes ranging from 6 years to 17 years at the 5% level and 8 to 11 years at the 10% level. This confirms that the observed trends have not arisen by chance, given that the observed cross-correlations are generally insignificant implying that the effective sample size is close to the actual sample size of 30.

In addition to the cross-correlation, the presence of serial correlation can complicate the identification of trends in that serial correlation can increase the expected number of false positive outcomes for the Mann-Kendall test. The magnitudes of the serial correlations calculated for the variables indicate that this is not a concern for the data analysed herein.

Although the observed trends occurred in all climatic regions investigated, differences between the results for the three climatic regions were observed. The YNBC has the largest percentage of trends and has trends that tend to be in the opposite direction from those in the other two climatic regions examined. This climatic region generally experiences increasing flow conditions, a tendency that is apparent in the low flow variable, Q7, the annual flow, and to a lesser extent, in the peak flows. This result is consistent with predictions from IPCC (1996) that northern latitude regions can be expected to experience increased runoff and increased flood frequencies. The less definitive peak flow response, in comparison to the other variables, may be partially a result of the nature of the peak flow variable that was selected. This work has used the annual daily peak flow as the measure of the high flow part of the hydrologic regime. It is possible that changes in the timing of the peak event, or changes in the number of flood events per year that exceed a given threshold, are occurring as a result of climatic changes. These types of changes would not be captured by the peak flow variable used herein.

The trends in the hydrologic variables for the YNBC climatic region appear to have a strong relationship with trends in the regional annual temperature series. Given that the regional temperature series has a strong increasing trend, and can be expected to increase further under global warming, the hydrologic trends identified for the catchments in this climatic region can be expected to continue in the future.

The NWF climatic region is characterized by reduced water availability. This is apparent in the low flow variable, the annual flow and the peak flow, all of which exhibit decreasing trends. This outcome is consistent with results from the work of Westmacott and Burn (1997) that investigated catchments from a similar area. There is again a strong connection between the regional annual temperature series for this climatic region and the hydrologic variables implying that the observed trends are associated with climatic change. The regional temperature series again exhibits a very strong increasing trend implying that the observed trends in the hydrologic variables represent a climatic change signal.

The catchments in the NEF climatic region experience similar impacts on the hydrologic variables to those noted for the catchments in the NWF climatic region. There is a weaker connection between the hydrologic responses noted and the regional annual temperature series for this climatic region. This could be anticipated because the regional temperature series does not exhibit a significant trend for the time period of interest.

The results of applying the trend detection technique to forested catchments can only be used to reach conclusions regarding the likely climate-change impacts on the specific catchments examined. The geographic similarity in the trend results may mean that it would also be safe to extrapolate the results to nearby catchments, with similar physiographic and climatic characteristics. However, the variability within the study area examined herein would imply that similar studies should be conducted for other geographic areas to determine the specific nature of the climate change impacts that are expected for catchments in other geographic settings.

It is not possible from the results presented herein to determine whether the trends identified are the result of natural climate variability or the result of actual change in the local climate. However, given the length of the data records analysed and the correlations noted between trends in hydrologic variables and trends in the temperature series, evidence points towards an actual change in the local climate as the cause of the trends.

Summary and Conclusions

This study has found a substantially greater number of trends in hydrologic variables than would be expected to occur by chance. The trends identified represent changes in the hydrologic regime that were correlated to changes in a regional temperature series. The nature and the direction of the identified trends are different for the different climatic regions examined. While increasing flow conditions are expected to occur in catchments in the Yukon/North BC Mountains climatic region, decreasing flow conditions are expected to occur in the North-East Forest and North-West Forest climatic regions. Further research is required to determine the expected climate change impacts on forested catchments on a larger geographic scale.

It is important to note that potential climate change is but one stress that forests experience. Forest perturbations, such as fire or harvesting, could exasperate the projected hydrologic impacts of climate change on forested catchments. The potential for climate change implies an increased need to manage our forest resources wisely.

Acknowledgements

The research described herein was supported in part by a grant from the Natural Sciences and Engineering Research Council of Canada (NSERC). This support is gratefully acknowledged.

References

Anderson, J.E., Shiau, S-Y. and Harvey, K.D. (1992) Preliminary investigation of trend/patterns in surface water characteristics and climate variations. In: Kite, G.W. and Harvey, K.D. (eds), *Using Hydrometric Data to Detect and Monitor Climatic Change. Proceedings of NHRI Workshop No. 8.* National Hydrology Research Institute, Saskatoon, SK, Canada, pp. 189-201.

Askew, A.J. (1987) Climate change and water resources. In: Solomon, S.I., Beran, M. and Hogg, W. (eds), *The Influence of Climate Change and Climatic Variability on the Hydrologic Regime and Water Resources.* IAHS Publication 168, pp. 421-430.

Burn, D.H. (1994a) Hydrologic effects of climatic change in west-central Canada. *Journal of Hydrology* 160, 53-70.

Burn, D.H. (1994b) The impact of climatic change on hydrologic variables. In: *Proceedings of 47th Annual Conference of the Canadian Water Resources Association, Winnipeg, Manitoba, Canada.* pp. 485-498.

Burn, D.H. (1994c) Identification of a data collection network for detecting climatic change. *Canadian Water Resources Journal* 19, 27-38.

Cleveland, W.S. (1979) Robust locally weighted regression and smoothing scatterplots. *Journal American Statistical Association* 74, 829-836.

Environment Canada (1999) *Establishment of the Reference Hydrometric Basin Network (RHBN) for Canada.* Atmospheric Monitoring and Water Survey Directorate, Environment Canada, Ottawa, Ontario.

Gan, T.Y. and Kwong, Y.T. (1992) Identification of warming trends in northern Alberta and southern Northwest Territories by the non-parametric Kendall's test. In: Kite, G.W. and Harvey, K.D. (eds), *Using Hydrometric Data to Detect and Monitor Climatic Change. Proceedings of NHRI Workshop No. 8.* National Hydrology Research Institute, Saskatoon, SK, Canada, pp. 43-56.

Gleick, P.H. (1986) Methods for evaluating the regional hydrologic impacts of global climatic changes. *Journal of Hydrology* 88, 97-116.

Gleick, P.H. (1987) The development and testing of a water balance model for climate impact assessment: Modeling the Sacramento Basin. *Water Resources Research* 26, 1049-1061.

Gleick, P.H. (1989) Climate change, hydrology, and water resources. *Reviews of Geophysics* 27, 329-344.

Gullet, D.W. (1992) Development of an historical Canadian climate database for temperature. In: Kite, G.W. and Harvey, K.D. (eds), *Using Hydrometric Data to Detect and Monitor Climatic Change. Proceedings of NHRI Workshop No. 8.* National Hydrology Research Institute, Saskatoon, SK, Canada, pp. 27-31.

Hirsch, R.M., Slack, J.R. and Smith, R.A. (1982) Techniques of trend analysis for monthly water quality data. *Water Resources Research* 18, 107-121.

IPCC (Intergovernmental Panel on Climate Change) (1996) *Climate Change 1995: The Second IPCC Scientific Assessment*. Houghton, J.T., Meira Filho, L.A. and Callendar, B.A. (eds). Cambridge University Press, Cambridge, U.K.

Kendall, M.G. (1975) *Rank Correlation Measures*. Charles Griffin, London.

Lawford, R.G. (1988) Towards a framework for research initiatives involving the impacts of climatic variability and change on water resources in the Canadian prairies. In: Magill, B.L. and Geddes, F. (eds), *The Impact of Climate Variability and Change on the Canadian Prairies: Symposium/Workshop Proceedings*. Alberta Environment, Edmonton, Alberta, Canada.

Lettenmaier, D.P. and Gan, T.Y. (1990) Hydrologic sensitivities of the Sacramento-San Joaquin River Basin, California, to global warming. *Water Resources Research* 26, 69-86.

Lettenmaier, D.P., Wood, E.F. and Wallis, J.R. (1994) Hydro-climatological trends in the continental United States 1948-1988. *Journal of Climate* 7, 586-607.

Lins, H.F. and Michaels, P.J. (1994) Increasing U.S. streamflow linked to greenhouse forcing. *EOS, Transactions, American Geophysical Union* 75, 284-285.

Livezey, R.E. and Chen, W.Y. (1983) Statistical field significance and its determination by Monte Carlo techniques. *Monthly Weather Review* 111, 46-59.

Mann, H.B. (1945) Non-parametric tests against trend. *Econometrica* 13, 245-259.

Pilon, P.J., Winkler, T., Harvey, K.D. and Kimmett, D.R. (1991) Hydrometric data in support of climate change studies in Canada. Presented at *NATO Advanced Research Workshop on Opportunities for Hydrological Data in Support of Climate Change Studies, Lahnstein, Germany, 26-30 August, 1991.*

Revelle, R.R. and Waggoner, P.E. (1983) Effects of a carbon dioxide-induced climatic change on water supplies in the western United States. In: *Changing Climate, Report of the Carbon Dioxide Assessment Committee*. National Academy Press, Washington, DC.

Running, S.W. and Nemani, R.R. (1991) Regional hydrologic and carbon balance responses of forests resulting from potential climate change. *Climatic Change* 19, 349-368.

Schadler, B. (1987) Long water balance time series in the upper basins of four important rivers in Europe – indicators for climate change? In: Solomon, S.I., Beran, M. and Hogg, W. (eds), *The Influence of Climate Change and Climatic Variability on the Hydrologic Regime and Water Resources*. IAHS Pubication 168, pp. 209-219.

Snedecor, G.W. and Cochran, W.G. (1980) *Statistical Methods*. Iowa State University Press, Ames, Iowa.

Westmacott, J.R. and Burn, D.H. (1997) Climate change effects on the hydrologic regime within the Churchill-Nelson River Basin. *Journal of Hydrology* 202, 263-279.

Chapter 4

The Potential Impact of Global Change on Surface Erosion from Forest Lands in Asia

K.F. Andrew Lo[1] and Qiangguo Cai[2]

[1]*Department of Natural Resources, Chinese Culture University, Taipei, Taiwan*

[2]*Institute of Geographical Sciences and Natural Resources, Chinese Academy of Sciences, Beijing, People's Republic of China*

The potential impacts of global change on forest cover throughout tropical Asia as well as influence of these changes on surface erosion by wind and water are discussed. Forests basically play two roles in the global warming phenomenon. First, forests have a passive role as the impact of global warming causes changes in their location, species mix, and growth. Second, forests play an active role as a vehicle for sequestering atmospheric carbon and thereby mitigating its buildup. Three examples presented from tropical forests in Malaysia, Indonesia, and Hawaii all indicate the importance of ground vegetation and soil litter cover (organic matter) in dissipating the erosive energy of rainfall and reducing overland flow. Temperature increases that would increase organic matter decomposition rates and reduce understorey plant cover would be projected to greatly increase the susceptibility of such sites to surface erosion. Agricultural conversion in hillslope forests with rugged topography, steep slopes and highly weathered and leached soils are especially susceptible to increased erosion hazards. Such hazards can be reduced by proper agro-management such as terracing, establishing a good legume cover, and protecting the exposed soil as early as possible.

Introduction

The most serious environmental problem facing the world today is the alteration of the composition of the atmosphere and the resultant climatic change. Various global-scale climatic models have been developed to simulate the climate of the planet. One of them is the Hadley Centre climatic model based on an increase in greenhouse gases according to the IPCC 'business-as-usual' emission scenario, without changes in sulphate aerosols (IPCC, 1990; UK Meteorological Office, 1998). Predictions indicate that high-latitude winters will warm faster due to feedback from the melting of sea-ice, and there will be some ocean areas where rises are quite limited. Rainfall changes will be most marked in tropical regions. The western Pacific region will experience temperature increases in the range from 0.8 to 2.4°C when the CO_2 concentration increases to 1.9 times of its pre-industrial level. Both summer and winter temperature will have the same degree of warming.

Rainfall scenarios based on global circulation models (GCMs) vary greatly, partly due to the complex processes controlling rainfall generation and partly due to the inherent temporal variability in rainfall (Loaiciga *et al.*, 1996; Hulme and Viner, 1998). In contrast to earlier estimates that precipitation changes would be minimal in subtropical latitudes (e.g., IPCC, 1990), Hulme and Viner (1998) developed one scenario for the extended tropics that exhibited the following seasonal rainfall patterns: (i) seasonal drying trends in southern and western Africa, central and western Australia and the Amazon Basin; (ii) seasonal increases in rainfall from tropical north Africa into India; and (iii) for Southeast Asia – December through February becomes drier, September through November becomes wetter, and little change occurs from March through August. Frequency of moderate to heavy storms was expected to increase in certain tropical areas, however, all rainfall scenarios need to be treated cautiously due to the limited length of the modelling period and the limitations of coarse-scale GCMs (Hulme and Viner, 1998).

Temperature variations were not uniform spatially. In recent decades, slight temperature decrease has been observed in the eastern part of the United States, south central Europe and eastern China, particularly during summer months (Kiehl and Briegleb, 1993). It has also been noted that from 1951 to 1990, the average maximum temperature during the day increased by 0.28°C, whereas the average minimum temperature at night increased by as much as 0.84°C (Harvey, 1995).

The projected changes in climate and increases in CO_2 could have significant direct and indirect impacts on agricultural systems, including effects on crop growth and yield, soil organic carbon and soil erosion (Crosson, 1993). As temperature rises, the growing season in high latitudes will be longer. However, daytime temperatures would at times exceed optimal level for crop growth, and increased evaporation rates would cause moisture stress in most plants. Smit (1987) estimates that the overall impact of rising temperature in Ontario, Canada would tend to be negative. In mid-latitude continental regions, marked decrease of soil moisture will have a negative effect on crop production. The global distribution of natural vegetation will also be affected by the greenhouse effect. Boreal forest will advance to higher latitudes and C_4 plants would be widespread (Cole and Monger, 1994). Plant species in high mountains will be pushed upward in elevation and may be eliminated if already growing near mountain summits. Based on the IPCC climate scenario, tropical forests will die back in many areas of northern Brazil. Additionally, increased emissions of CO_2 and trace gasses from burning and clearing of tropical forests would exacerbate IPCC global warming estimates (Fearnside, 2000). Many tropical types of grassland will be transformed to desert or temperate grasslands. On the Tibetan Plateau, a 2°C increase in annual temperature would cause most of the current ecosystems to disappear and in the central and northern sections, be replaced with desert (Brown, 1995).

Changes in climate and CO_2 can alter the rates of soil erosion by wind or water. Soil erosion can be directly affected by changes in erosive forces, for instance, by the amount and intensity of precipitation (water erosion) and wind speed and direction (wind erosion) during individual events. Changes in climate and CO_2 can indirectly affect erosion by influencing the degree and timing of crop cover and the production and decomposition of residue. Changes in soil water content, as affected by changes in the ratio of precipitation to evapotranspiration, can also influence erosion. Generally, water erosion increases and wind erosion decreases as soil becomes wetter (Wischmeier and Smith, 1978). Water and wind erosion both tend to be dominated by extreme events.

Throughout the world mankind has developed, even created lands that produce a bounty of agricultural products. At the same time, as a result of human activities, soil productivity (the capacity of the soil to produce, quantitatively and/or qualitatively, goods or services) is often decreased at rates many times greater than would occur naturally. The processes that individually or collectively reduce soil productivity are called land degradation processes. According to FAO (1979), land degradation processes include water erosion, wind erosion, salinization, sodication, chemical, physical as well as biological degradation. Among them, both water and wind erosion are considered the most important forms of land degradation. Although there appears to be little public awareness of land degradation as an environmental problem compared to deforestation, desertification or acid precipitation, scientists agree that land degradation is one of our most serious problems (Brown, 1984; Lal, 1988).

In this chapter, we discuss the potential impacts of global change on forest cover throughout tropical Asia as well as the influence of these potential changes on surface erosion by wind and water. Forest cover protects the underlying soil from the direct effects of rainfall. Overland flow is generally low from undisturbed forest lands in tropical Asia (e.g., Bonell, 1993; Noguchi *et al.*, 1997), however, localized overland flow may occur during very high intensity rains. Tree roots bind the soil and the litter layer protects the ground from rainsplash. With the die back of forest, rates of soil loss will rise and mass movements will increase in magnitude and frequency.

Factors Influencing Soil Erosion on Forest Lands

The combination of raindrop size, velocity and shape, storm duration, and wind speed control the erosive power of rainfall. Well-managed forests are unsurpassed as a vegetative cover for watersheds. Leaves, branches, and the organic layer on the forest floor break the velocity of falling raindrops (Lowdermilk, 1930). Rain may slowly infiltrate or be held by the spongy leaf litter and fine root matrix. Some water moves down branches and trunks and infiltrates into the soil along root channels. The net result is that a properly managed forest has higher infiltration and less runoff, erosion, and sedimentation than it would have as cropland (Dohrenwend, 1977). Runoff and erosion from well-maintained forestland are small, often less than 5% and 1%, respectively, of runoff and erosion from bare soil (Bennett, 1939). Furthermore, slow, deep seepage gives rise to many springs and delays and reduces flood peaks.

Raindrops falling from the canopy are often larger than the original raindrops because they coalesce on the canopy (Stocking and Elwell, 1976). Typical total kinetic energies of throughfall have been found to be 4–50% greater than those experienced in open conditions (Wiersum, 1985; Vis, 1986). However, under natural forests this process of drop formation may be mediated by the subsequent interaction with multiple canopy levels, resulting in droplet break-up on impact (Vis, 1986) and, closer to the ground surface, prevention of the attainment of terminal velocities (Brandt, 1988). The litter layer also significantly dissipates raindrop energy. The water reaching the mineral soil remains clear, and the soil pores at the surface are not infilled with detached soil particles.

If rainfall intensities below the forest canopy exceed the infiltration capacity of the soil, the unabsorbed water (minus any losses from evaporation of surface water stores and surface detention storage) moves down slope via the hydrological process known as infiltration-excess overland flow. Generally, however, infiltration rates are usually high on the forest floor as a consequence of the high permeability of the forest litter layer, good soil aggregate structure and the presence of macropore channels formed by roots and soil fauna activities (Lal, 1987; Bonell, 1993). Thus in natural forests, infiltration-excess overland flow is a relatively limited phenomenon, only occurring where litter dynamics produce ephemeral patches of bare ground (Spencer *et al.*, 1990); where soil becomes exposed through tree fall; and where, on steep slopes during intense storms or earthquakes, landslides expose the regolith (Garwood *et al.*, 1979). Elsewhere, surface wash, and associated soil losses, are usually insignificant.

With prolonged, heavy tropical rains, a perched water table may rise to the surface resulting in saturated overland flow. If surface runoff does occur, the leaves and roots of plants inhibit movement of soil particles. Leaves form a rough surface, impede flow, and reduce the velocity of flowing water; roots act to bind the soil (Dyrness, 1967; Hashim, 1988).

When a forest is disturbed by trampling of livestock or by logging operations, the soil is compacted, thus increasing its bulk density and soil strength, decreasing water infiltration (Froehlich, 1988). The natural protection from erosion may also be destroyed. Extensive tree removal reduces transpiration and may leave the subsoil perennially wet and impervious. Also, the sun reaching the soil surface causes rapid decomposition of organic layers. With the protection from raindrop impact taken away, the soil begins to erode (Tsukamoto, 1975).

When surface cover is removed, raindrop impact may change surface aggregate structure, loosening fine soil fractions, blocking macropores and perhaps also micropores (Lawson *et al.*, 1981), and increasing soil erodibility (Luk, 1979). Reduction in canopy cover and litter cover can therefore result in increased soil losses with the litter interface between the atmosphere and soil being the most critical factor (Wiersum, 1985).

The most effective way to reduce wind erosion is to cover the soil with a protective mantle of growing trees or with a thick mulch of tree residue. Barriers of plant material increase the thickness of the blanket of still air next to the soil (Lyles and Tatarko, 1986). The protection that forest cover provides is influenced by tree species (amount of vegetative cover and time of year when cover is provided), tree geometry and density, and row orientation (Woodruff and Siddoway, 1965). Litter residues left on the surface, especially if tall and dense, offer almost as much protection as growing trees (Lyles, 1988). A complete cover of growing trees offers maximum protection, but individual trees and rows of trees across the direction of the wind also reduce ground-level wind velocity and erosion. Tree barriers are effective because air is a fluid. As air moves up a porous barrier, part of it is pushed over or around the barrier. Air that is not deflected passes through the barrier at a fast rate (funnelling effect), then immediately slows down as it spreads out to occupy all the space behind the barrier (Hagen, 1976). Wind speed returns to normal only when the deflected air returns to its initial position in the wind stream.

Impact of Climatic Change on Surface Erosion

Atmospheric temperature has changed continuously throughout the Earth's history. Some changes are small while others are large. Natural factors cause some of these temperature changes, but other changes are related to anthropogenic effects. With the presence of mankind, especially after the industrial revolution of the 19th century, human influence on weather has overwhelmed the natural climatic variation, especially during the last 100 years. Chiang and Lo (1989) studied the annual mean air temperature between 1890 and 1988 for nine stations in Taiwan at Pengchiyu, Keelung, Taipei, Taichung, Hualien, Penghu, Tainan, Taitung and Hengchun. Results indicate an increasing mean annual temperature trend (Table 4.1). The increase is about 0.7°C for the last 90 years. Increases during the summer months are greater than 0.9°C.

During the winter months these increases are less significant. The interannual temperature difference is larger in the central and the southern part of Taiwan.

Table 4.1. Temperature warming trend in Taiwan.

Station	Mean annual temperature increase for the last 92 years (°C)
Taipei	0.75
Keelung	0.54
Pengchiyu	0.79
Taichung	0.64
Hualien	0.71
Penghu	0.68
Tainan	0.91
Taitung	0.72
Hengchun	0.69

Source: Chiang and Lo, 1989

Gornitz and Lebedeff (1987) suggested that global warming would lead to accelerated evaporation of seawater and increased rainfall and stream runoff. However, at regional scales, rainfall and runoff may vary considerably (Loaiciga *et al.*, 1996; Hulme and Viner, 1998). With a ground temperature rise of 0.5°C, IPCC model results indicate a 10% increase in rainfall (UK Meteorological Office, 1998). Rainfall increases will occur both at high and low latitudes in the northern hemisphere. Rainfall at the mid latitudes, on the other hand, will decrease. If model results are accurate, most areas in the northern hemisphere will have substantially less water stored in the soil and annual stream runoff will be lower.

Chiang (1995) found a decreasing trend in monthly rainfall data in Taiwan from 1937 to 1994. The linear regression relationship is

$$Y = 2588.6 - 1.71\ X$$

where, Y is the annual rainfall amount (mm) and X = Year – 1911.

Using the most recent 20 years (1977–1994) of rainfall record, Lo (1999) re-evaluated the rainfall erosivity factor according to the Universal Soil Loss Equation in Taiwan. By adding 20 years of data to the original record (1957–1976), the accuracy and applicability of the rainfall erosivity factor is improved. Results show that the rainfall erosivity factor for most of the precipitation stations in Taiwan exhibited a slight to moderate decrease as a result of adding 20 more years' data to the long-term record (Table 4.2).

Rainfall and rainfall erosivity reductions may lead to decreases in annual runoff. WRPC, DEA (1995) predicts that by 2050 annual runoff in Taiwan will decrease 4%. For extremely dry years, the annual runoff will decline by more than 4.4%. However, surface soil erosion may react differently to rainfall variations. During summer months, the prolonged dry interval between two storms would cause a deficit of water in the soil and moisture available for plant growth. Such climatic change would lower the groundwater level and stream base flow.

Table 4.2. Change in the rainfall erosivity value based on longer rainfall records, represented as percentage of rainfall stations to the total number of rainfall stations.

Rainfall erosivity region	Station with higher erosivity value (%)	Station with lower erosivity value (%)
Taipei and northwestern coast	73.9	26.1
Northeast	22.2	77.8
Taichung and western coast	21.7	78.3
Northern and south-central mountain	41.2	58.8
Southwestern mountain and coast	21.9	78.1
Pingtung	20.0	80.0
Taitung	11.1	88.9
Hualien	54.8	45.2

Source: Lo, 1999

Changes in rainfall will be associated with changes in temperature because cloud cover affects the temperature below clouds and water in clouds tends to absorb heat slowly and release it again slowly (Lau and Yang, 1996). Where rainfall increases the weather may be generally cooler, but in areas that become drier extremes of temperature may be more common. Extremely hot, dry summers and extremely cold winters may occur more often.

Soil temperature and soil water are two important crop production factors. They would change along with climatic change. Any climatic perturbation would exert immediate impact on crop production (Ausubel and Biswas, 1980). Soil temperature affects the type of crops grown and the amount of ground cover that exists. The same amount of precipitation is more effective in producing ground cover where temperatures are cool than where they are very warm. This is because cool temperatures increase moisture effectiveness due to less evaporation and transpiration. Precipitation that would produce a semiarid climate in a cool region may produce an arid climate in a warmer one, depending on soil texture (e.g., vegetation growing on sandy soils utilizes limited amounts of spring and summer rains much more effectively than on clay or loam soils).

Rainfall has a direct effect on the amount of ground cover produced, which in turn influences the extent of protection against wind and water erosion. The effectiveness of rainfall has also, in the past, determined the type of soil that has formed and the amount of organic matter present in the surface horizon (Robinson *et al.*, 1996). Soils in the more arid regions are generally low in organic matter because warm temperatures have resulted in more rapid decomposition. The lower organic matter content makes the soil more susceptible to erosion during intense rains that occur in such regions.

Temperature variation has an effect on erosion similar to that between climatic regions, but on a smaller scale. With higher temperatures, the soil becomes drier. Consequently, soils are lower in organic matter and are more likely to have sparse vegetative covers in areas of limited rainfall. The aggregates are more likely to be very dry and susceptible to rapid dispersion during rainfall events (Hudson, 1981).

Temperature changes also affect infiltration and percolation by their influence on the viscosity of water. The higher the temperature, the lower the viscosity of water and the higher rate of infiltration and percolation. The actual effect of temperature change, however, is quite small compared to that of other factors unless the change is over a wide range (Smith and Wischmeier, 1962).

The warming effect will not be spatially uniform. Temperature increase at the poles is likely to be two to three times the global average, while increase in the tropics may be only 50 to 100% of the average. This increased warming at the poles will reduce the thermal gradient between the equator and the high latitude regions, decreasing the energy available to the "heat-engine" that drives the global weather machine.

As the thermal gradient is reduced, global patterns of winds and ocean currents, as well as the timing and distribution of rainfall, will change. This situation may result in less serious wind erosion in many parts of the world. However, there are areas where wind erosion is just as environmentally degrading as the worst water erosion. Wind erosion is the process by which loose surface material is picked up and transported by the wind, and surface material is abraded by windborne particles. The spatial redistribution and resorting of particles by wind erosion may profoundly affect soils, related microtopography, and any agricultural or forestry activity associated with the deposits (Chepil, 1950; Zobeck, 1991). Factors that influence the likelihood of wind erosion include soil conditions, rainfall, and vegetation characteristics.

The physical nature of the soil will determine the ease with which particles are dislodged, but a far more important consideration is that only dry soil becomes airborne; moist soil will not be moved by wind (Chepil, 1956). Antecedent soil moisture, therefore, has a large influence on wind erosion, particularly the extent to which the soil will be dried by hot winds and low humidity, and also the rainfall regime. Wind erosion is more frequent when the mean annual rainfall is low, and in regions with higher rainfall but long dry periods (Bisal and Hsieh, 1966). Wind erosion of soils is negligible when a continuous vegetation cover exists. Wind erosion is particularly significant in both hot and cold deserts, coastal dune areas and exposed mountain regions (Bagnold, 1937, 1973). The consequences of human activity and erosion are most serious in those agricultural areas that experience low, variable and unpredictable rainfall, high temperatures and rates of evaporation, and high wind speed, as is the case in semiarid areas.

Case Studies of Existing Forest Degradation

The effects of forest vegetation on soil and water conservation have been well documented in tropical regions. The following are several examples to illustrate these effects.

West Java, Indonesia

Wiersum (1985) showed that Acacia (*Acacia auriculiformis*) tree canopies decreased rainfall water reaching the soil by 11.8%, however, the erosive power of the rainfall increased by 24.2%. Litter reduced the erosive power by as much as 9.5% in comparison with bare soil, and the presence of undergrowth decreased erosion by another 3.7%. Although these data refer to different aspects of the erosion process and cannot be compared directly, they do suggest that litter and understory vegetation protects the soil to a greater extent than such non-indigenous plantation canopies.

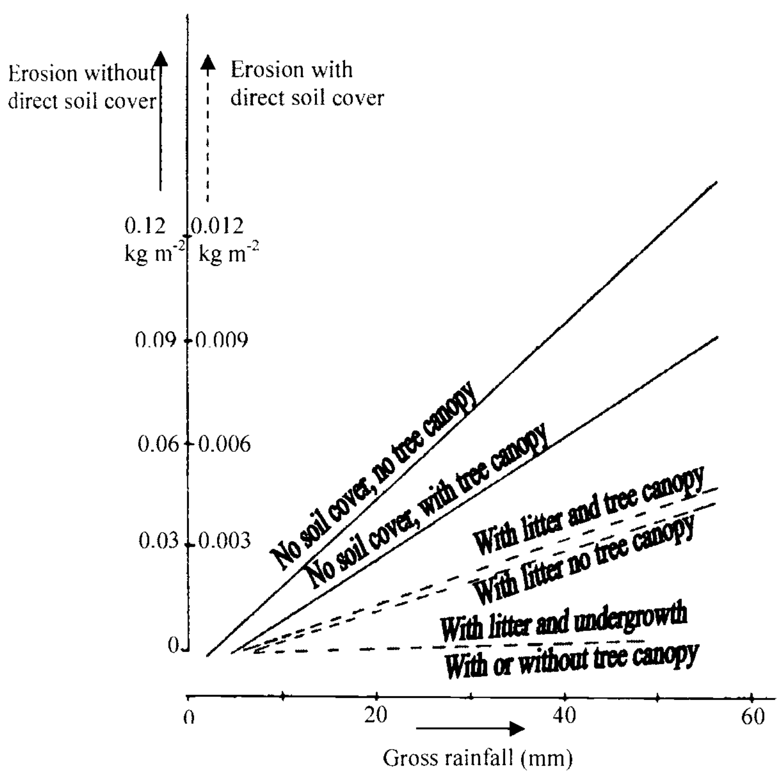


Fig. 4.1. Theoretical relations between erosion and 24-hour rainfall, depending upon the presence of various vegetation layers (Wiersum, 1985).

The importance of a direct litter cover over the mineral soil as a protection mechanism against surface erosion is demonstrated in Fig. 4.1. The sustained presence of litter is ensured by the litter production capacity of the tree canopies. The litter decomposes gradually, resulting in increased humus in forest soils and decreased erodibility. According to the IPCC climate scenario, the predicted temperature change will have a negative effect on vegetative cover. Declining ground vegetation will lead to more severe surface erosion as shown in Fig. 4.1.

Kemaman, Malaysia

Although trees provided an aerial vegetation cover, substantial soil erosion occurred under tree crops when compared to natural forest conditions (Table 4.3) (Hashim, 1988). Much less surface wash will be likely to occur during intense storms at sites with live ground vegetation compared to sites covered only by leaf litter. Live ground vegetation prevents soil detachment and transport much more effectively than only leaf litter. There are indications that a high level of soil conservation can be achieved in a system where both trees and dense live ground vegetation are present. As temperatures increase and soils become drier, live ground vegetation will decline. Soil detachment and transport will occur more readily. Under trees where there is no live ground vegetation, the runoff is relatively less impeded and scours the soil surface more readily. Rainfall intercepted by the leaves and branches reaches the forest floor as canopy drip and stem flow (Spencer *et al.*, 1990). These processes play significant roles in soil detachment and transport during intense storms.

Table 4.3. Soil loss and runoff under different vegetative covers for the period January-December 1986 in Malaysia.

Treatment	Runoff (mm)	Soil loss (t ha^{-1})
Dense ground vegetation	45.5	5.76
Dense ground vegetation + Gliricidia	6.3	0.28
Cocoa + Gliricidia	76.4	1.92
Forest	4.5	0.15

Source: Hashim, 1988

Hilo, Hawaii

With the maturation of Eucalyptus (*Eucalyptus saligna*) stands, runoff decreased when litter was present on plots (Table 4.4) (Lo and El-Swaify, 1987). This decrease may have been caused by three mechanisms. First, litter and humus accumulate as the stand matures and, in turn, increases the water absorption capacity of the soil. Second, litter reduces overland flow downslope allowing more infiltration. Finally, the presence of litter prevents crust formation and therefore maintains the infiltration capacity of the soil at a high level.

Table 4.4. Effect of tree litter on soil erosion in Hawaii (monitoring period from 1983 to 1986).

Treatment	Runoff (mm)	Soil loss (t ha^{-1})
Eucalyptus + tree litter	436.04	0.24
Eucalyptus – tree litter	774.26	16.96
Bare	891.47	2.85

Source: Lo and El-Swaify, 1987

Effects of Forest Cover Change on Erosion and Sedimentation

With the removal of litter, either by human intervention or as a result of climate change and forest dieback, the rate of soil loss increases as the stand matures. This is attributed to an increase in the percentage of throughfall reaching the soil surface, as highly erosive water drops are concentrated onto a small fraction of the surface due to the shading effect of tree canopies. This effect is larger for older trees since canopy drip occurs from higher elevations. Interestingly, plots maintained bare throughout stand development experienced less soil erosion than tree-covered but litter-less plots. It was concluded that the primary attribute of forest vegetation is the maintenance of an effective and protective ground cover at the immediate soil surface.

In undisturbed or mature secondary forest ecosystems, water movement under saturated conditions takes place in soils through macropores that dominate the active pore space and, therefore, surface runoff is generally low (Lawson *et al.*, 1981; Sidle *et al.*, 2000). The removal of the vegetation cover from the soil generally results in an increase in bulk density, decrease in porosity, and a reduction in infiltration capacity depending on the level of site disturbance and compaction associated with logging operations or site conversion practices (Wood, 1977). A rapid decline in fertility has also been reported following vegetation removal (Florence, 1967). Fig. 4.2 gives a generalized picture of the different ecological and management phases in the conversion of natural forests to short-rotation plantation. Effects on infiltration for various ecological and vegetation conditions in India are given in Table 4.5.

As previously mentioned, the presence of trees in the landscape is generally beneficial to the hydrological characteristics of the site. Studies by Pereira (1973) and Patnaik (1978) have demonstrated the favourable influence of trees on the hydrological characteristics and water balance. The development of a tea plantation in a tall rainforest caused a considerable increase in storm runoff during the initial clearing and terracing operations, although the tea canopy, once established, gave good protection (Pereira *et al.*, 1962). In addition to these direct effects of trees on the physical properties of soils, there are also indirect advantages that could benefit the forest ecosystem. For example, one of the well-accepted effects of mulching is the improvement of soil physical conditions. Several reports indicate that better soil physical conditions enhance better vegetation growth (Lal and Greenland, 1979).

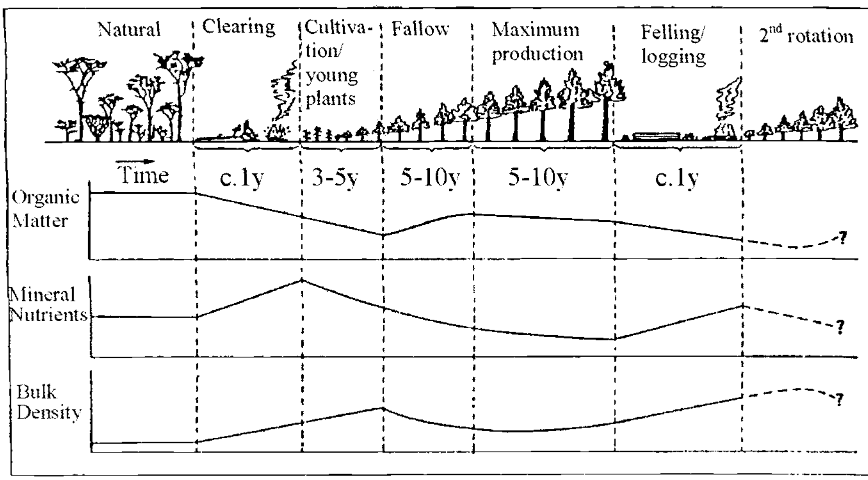


Fig. 4.2. A generalized picture of the likely changes in organic matter status, nutrient content and bulk density of topsoil during different management phases in the natural forest to short-rotation plantations (Lundgren, 1978).

Table 4.5. Infiltration capacity under different vegetative cover types in India.

Location and vegetative cover/land use	Infiltration capacity (mm h^{-1})
Ootacamund Study No. 1	
Shola forest	168
Blue gum plantation	207
Grazed grassland	51
Ranchi	
Forest	200
Permanent natural grass	80
Cultivated farmland	20

Source: Tejwani, 1979

The protective role of trees in imparting stability to the whole ecosystem is also very well known. Clearing vegetation affects not only the farmlands in the immediate vicinity (on-site), but also generates off-site impacts such as flooding, pollution of rivers and rapid silting of dams. A study of 17 reservoirs in India showed that when trees were removed in the catchment areas, the annual sedimentation rate was 917 m^3 km^{-2}, resulting in a 67% decrease in the expected life of the dams (Tejwani, 1977). Consequently, the incidence of flooding was much higher. The increased flooding was partly attributed to the denudation of catchment areas. Flood damage below the deforested Himalayan catchment areas has resulted in crop losses of almost US$250 million per annum and has affected an average of 6 million ha per year from 1971 to 1980 (World Bank/FAO, 1981).

Forests and other tree plantations may also influence regional microclimate. Because the albedo in forests is low, some researchers have claimed that an increase in albedo can cause a decrease in rainfall (Charney *et al.*, 1975). Thus, if plant cover diminishes, a decrease in rainfall may occur that could initiate or aggravate drought. With their thick and complex structure, forests have low heat conductivities and large amounts of latent heat are involved in the evapotranspiration process. This enables the forest to act as a buffer against rapid cooling or heating and thus regulates the heat released to the ground. Nair and Balakrishan (1977) observed markedly reduced rates of evaporation and lower temperature regimes even within a plantation of coconut palms (which have long unbranched stems) compared to the bare soil (Table 4.6). These favourable attributes of trees and their significant influence on soil conservation, physical soil properties, the hydrological balance, and microclimate effects, attest to their benefits in conserving and stabilizing ecosystems. With the dieback of trees due to climate change, current ecosystems may change or even disappear, resulting in increased surface erosion and land degradation.

Table 4.6. Evaporation from different microclimates of coconut-based systems expressed as percentages of evaporation from the open area on the west coast of India.

Month	Open area	Unirrigated coconut	Irrigated coconut	Coconut + cacao
November	100 (7.4)*	75.7	66.2	36.5
December	100 (6.2)	75.8	69.4	30.6
January	100 (5.8)	79.3	62.1	27.6
February	100 (5.3)	83.0	56.6	24.5
March	100 (5.0)	102.0	76.0	32.0
April	100 (6.1)	86.9	68.8	30.6
May	100 (5.8)	84.5	58.6	29.3
Mean	100	80.9	65.4	30.1

Source: Nair and Balakrishnan, 1977. *Figures in parentheses are the mean monthly values of evaporation (mm day^{-1}) from the open area.

Forested land has been cleared for centuries to make way for farms. This seems natural, for most of the European forests were cleared to provide farmland – not, as some people suppose, to provide timber for construction or fuel for industry. Although tropical soils are not well suited to agriculture, the climate can be quite desirable and with care the soils can be managed.

However, farming may be of little benefit to local people. Much of the current conversion of forest lands in tropical Asia is currently practised in steep terrain.

Lo and Chiang (1996) conducted a study in a reservoir watershed in Taiwan to evaluate the impact of various types of land conversion on the sediment and nutrient yields. They adapted the AGNPS (Agricultural Non-point Source Pollution) model for this study. Model input parameters were initially modified to reflect several large-scale land use change scenarios. These included transforming existing land areas with slope steepness less than 30% to orchards or forest land, and converting slopes less than 50% to orchards or forest land. Model outputs are summarized in Table 4.7 for each type of land use change. Obviously, increasing forest land in a reservoir watershed tends to decrease the runoff volume, sediment yield, erosion, and nutrient losses. Converting more areas to fruit plantations, on the contrary, results in substantial increases in sediment production, erosion, and sediment-associated nutrient losses. However, runoff volumes as well as water-soluble nutrients increase only slightly. This is in part due to a relatively small runoff coefficient difference between forested land and orchards, and the strong dependence of water-soluble nutrient yields on runoff volume.

Table 4.7. Impact of alternative land use schemes on sediment and nutrient yields.

AGNPS parameter (unit)	Original watershed condition	Land converted to orchard		Land converted to forest	
		<30% slope	<50% slope	<30% slope	<50% slope
Runoff volume (mm)	1872	1869	1867	1755	1588
Soil erosion (t ha^{-1})	117.6	124.1	156.1	117.5	118.6
Sediment (10^6 t)	1.87	2.08	2.68	1.69	1.61
Sediment					
N (kg ha^{-1})	48.1	52.4	64.1	44.3	42.7
P (kg ha^{-1})	24.0	26.2	32.0	22.2	21.3
Soluble					
N (kg ha^{-1})	15.0	15.0	15.1	14.2	12.9
P (kg ha^{-1})	0.9	0.9	1.0	0.9	0.8
COD(kg ha^{-1})	934.6	1387.2	2035.0	876.0	793.3

Source: Lo and Chiang, 1996

In order to estimate the potential benefits of maintaining forest cover, the model inputs were artificially altered to have 100% (entire watershed covered with forest), 81% (forested land <30% slope converted to orchards), 53% (forested land <50% slope converted to orchards) and 0% (entire watershed converted to orchards) forest cover (Lo and Chiang, 1996). Model outputs are shown in Table 4.8. Results indicate that sediment, erosion, runoff, nutrient and COD (chemical oxygen demand) levels decrease with increased forest cover. Between 0 and 100% forest cover, runoff declines 30%; erosion decreases by 55%; and total sediment decreases 69%. As for the nutrient flow, soluble N and P decline 30 to 40%; sediment N and P decline 60 to 61%; and soluble COD decreases more than 80%. Therefore, maintaining good forest cover significantly reduces soil movement, total sediment, nutrients, and soluble COD at the watershed outlet. Similar results (Table 4.9) have been reported in the granite slope erosion areas of Guangdong, South China (Chang *et al.*, 1994).

Table 4.8. Potential benefits of maintaining forest cover in the reservoir watershed.

AGNPS output		Per cent forest cover			
Parameter	Unit	0	53	81	100
Runoff volume	mm	1867	1580	1415	1300
Soil erosion	$t\ ha^{-1}$	306.7	175.5	144.6	138.0
Sediment	10^6 t	5.33	2.75	2.07	1.65
Sediment					
N	$kg\ ha^{-1}$	111.0	65.4	52.2	43.4
P	$kg\ ha^{-1}$	55.5	32.7	26.1	21.7
Soluble					
N	$kg\ ha^{-1}$	15.1	12.8	11.4	10.6
P	$kg\ ha^{-1}$	1.0	0.8	0.7	0.6
COD	$kg\ ha^{-1}$	3169.1	1891.1	1160.5	649.3

Source: Lo and Chiang, 1996

Table 4.9. Relationship of soil erosion area and forest cover in Guangdong Province, South China.

County	Land area (km^2)	Forest cover percentage	Soil erosion area (km^2)	Eroded land area (%)
Wu-hua	3218.24	16.40	875.83	27.21
Hsing-ling	2080.60	22.50	600.38	28.86
Ying-the	5745.20	27.43	1430.90	24.91
Hsin-hsing	1934.68	31.65	196.45	10.15
Da-po	2471.30	35.97	396.53	16.05
Yueh-chang	2387.00	43.01	386.60	16.20
Chu-jiang	3173.20	50.89	319.20	10.06
Ping-yuan	1381.80	54.30	178.26	12.90
Jiau-ling	962.75	55.00	53.54	5.56
Shih-hsing	2174.10	63.75	157.80	7.26
Guang-ling	1707.48	64.13	114.28	6.69
Jen-hua	1821.20	69.58	164.00	9.01

Source: Chang *et al.*, 1994

Deforestation can lead to reduced water infiltration into soils and consequently higher rates of runoff after rainfall depending on the type of logging or land conversion operations (e.g., Ballard, 1988). The reduced infiltration capacity results from increased bulk density due to soil compaction, soil structure degradation due to soil disturbance and exposure to raindrop energy, and surface sealing by detached particles (Rice *et al.*, 1972). Associated with large runoff volumes are large increases in soil loss. In some worse cases, large sections of the hillsides may slip downslope causing massive landslides (e.g., Sidle *et al.*, 1985). Lee (1981) reported that the most serious types of erosion in Taiwan are landslides and stream bank sloughing. A total of more than 7810 areas of landsliding have been reported, covering areas of more than 10,000 ha. Resultant damage includes loss of cropland (both from topsoil stripping and from burial), property damage including collapse of buildings and breaching or filling of engineering earthworks, covering or undermining of roads and culverts, destruction of timber, and loss of human life. Sediments delivered with runoff water are also harmful to aquaculture and marine ecosystems. Several studies have shown that most of the nitrogen and more than 90% of the phosphorus moving from fields to streams is carried by sediment (Crathorne and Dobbs, 1990; Carter and Turnock, 1993). These effects are determined by both quantity and quality of sediments. In general, sediments derived from highly weathered tropical soils cause more turbid runoff compared to those derived from temperate soils (El-Swaify, 1987).

Table 4.10. Siltation rates in selected reservoirs.

Country	Reservoir	Annual siltation rate (t)	Years to fill with silt
Egypt	Aswan High Dam	139,000,000	100
Pakistan	Mangla	3,700,000	75
Phillipines	Ambuklao	5,800	32
Tanzania	Matumbulu	19,800	30
Tanzania	Kisongo	3,400	15

Source: El-Swaify *et al.*, 1982

Soil transported from farms by runoff may end up in local streams, rivers, canals, or irrigation and hydroelectric reservoirs. The loss of topsoil that reduces land productivity may also reduce irrigation water quality, electrical power generation, and navigability of waterways. Table 4.10 shows an example of a typical situation in Pakistan. The designers of the Mangla Reservoir projected a life expectancy for the dam of at least a century. However, they neglected the effects of an increasing population in the contributing watershed. Progressive land clearing and cultivation in the headwaters of the drainage leads to a rate of siltation that will probably fill the reservoir at least 25 years earlier than projected (Eckholm, 1976). In the Philippines, scores of hydroelectric and irrigation reservoirs have been constructed with assistance from international development agencies. As in Pakistan, the combination of forest conversion and steep slopes being cleared for cultivation has generated record siltation rates (El-Swaify *et al.*, 1982). In Taiwan, streams often transport tremendous amounts of sediment from upper reaches to downstream areas and reservoirs due to inappropriate land use in upper portions of watersheds (Chiang, 1995; Lo and Chiang, 1996). This results in an aggradation of the streambed and decreases the capacity of reservoirs (Table 4.11). For instance, the Shihmen Reservoir, one of the largest reservoirs in Taiwan with a capacity of 200 million m^3, received a total of 19 million m^3 of sediment within 1 year of its completion (Lee, 1981).

Table 4.11. Sedimentation rates measured at selected reservoirs in Taiwan.

Reservoir	Catchment area (km^2)	Annual deposition ($m^3\ km^{-2}$)
Shihmen	763.4	3,789
Tsengwen	481.0	13,508
Wu-shih	204.8	7,018
A-kun-tien	31.9	13,350
Paiho	26.6	19,853
Wu-shan-tou	60.0	22,970
Mingteh	61.1	7,226

Source: Lee, 1981

Conclusions

Trees intercept falling rain, thus protecting tropical soils from erosion and compaction while stabilizing soil via their strong and dense root matrix, preventing mass soil movement. The buffering effects of tree canopies on raindrop impact is especially valuable in the tropics where large raindrops fall rapidly with high energies due to severe storms. Trees contribute organic matter to the forest floor where the spongy litter layer of tropical forests absorbs raindrop impact. This litter also slows the flow of water across the rainforest floor. Trees also perform major ecological functions such as providing food and cover for a variety of tropical flora and fauna.

Forests basically play two roles in the global warming phenomenon. First, forests have a passive role as the impact of global warming causes changes in their location, species mix and growth. Second, forests play an active role as a vehicle for sequestering atmospheric carbon and thereby mitigating its buildup. The land area of the forests required to stabilize the level of atmospheric CO_2 at current levels is estimated to be a minimum of 465 million ha. The economic cost of establishing tree plantations of this size is estimated to be at least US$372 billion (Sedjo, 1989). With the dieback of trees due to climate change, current ecosystems will change and even disappear, resulting in increasing surface erosion and land degradation. The need for proper and effective forest management and conservation in natural forests is well recognized for controlling soil erosion (Brown, 1993). Hillslope forests with rugged topography, steep slopes and highly weathered and leached soils are highly susceptible to increased erosion hazards. Such hazards can be reduced by proper agro-management such as terracing, establishing a good legume cover, and protecting the exposed soil as early as possible (Jorgensen and Craig, 1983).

References

Ausubel, J. and Biswas, A.K. (1980) *Climatic Constraints and Human Activities*. IASA Proceeding Series, Volume 10, Pergamon Press, New York.

Bagnold, R.A. (1937) The transport of sand by wind. *Geographical Journal* 89, 409-438.

Bagnold, R.A. (1973) *The Physics of Blown Sand and Desert Dunes*. Chapman & Hall, London.

Ballard, T.M. (1988) Soil degradation associated with forest site preparation. In: *Degradation of Forested Land: Forest Soils at Risk. Proceedings 10th B.C. Soil Science Workshop: 3-12, Feb. 1986*. Land Mgmt. Rep. 56. Ministry of Forests, British Columbia, Canada, pp. 74-81.

Bennett, H.H. (1939) *Soil Conservation*. McGraw-Hill, New York and London.

Bisal, F. and Hsieh, J. (1966) Influence of moisture on erodibility of soil by wind. *Soil Science* 102, 143-146.

Bonell, M. (1993) Progress in the understanding of runoff generation dynamics in forests. *Journal of Hydrology* 150, 217-275.

Brandt, J. (1988) The transformation of rainfall energy by a tropical rain forest canopy in relation to soil erosion. *Journal of Biogeography* 15, 41-48.

Brown, L.R. (1984) The global loss of topsoil. *Journal of Soil and Water Conservation* 39, 162-165.

Brown, L.R. (1995) *State of the World 1995*. A Worldwatch Institute Report, W.W. Norton & Co., New York.

Brown, S. (1993) Tropical forests and the global carbon cycle: The need for sustainable land use patterns. *Agriculture Ecosystems and Environment* 46, 31-44.

Carter, F.W. and Turnock, D. (1993) Problems of the pollution scenario. In: Carter, F.W. and Turnock, D. (eds), *Environmental Problems in Eastern Europe*. Routledge, London, chapter 9.

Chang, K.L., Yuang, C.Y. and Li, W.S. (eds) (1994) *Studies of Hua-Nan Slope Lands*. Science Publishing, Beijing, China (in Chinese).

Charney, J., Stone, P. and Quirk, W. (1975) Drought in the Sahara: Biogeophysical feedback mechanism. *Science* 187, 434-435.

Chepil, W.S. (1950) Properties of soil which influence wind erosion: I. The governing principle of surface roughness. *Soil Science* 69, 149-162.

Chepil, W.S. (1956) Influence of moisture on erodibility of soil by wind. *Soil Science Society of America Proceedings* 20, 288-292.

Chiang, S.H. (1995) The threat of water resources in Taiwan. In: *Proceedings of Symposium on Regional Climatic Changes Causes and Effect, Taipei, Taiwan*. pp. 90-103 (in Chinese).

Chiang, S.H. and Lo, K.F.A. (1989) Temperature warming trends in Taiwan. *Proceedings of Workshop on Long-term Air Changes and Their Climatic Impact, Taipei, Taiwan*. pp. 6-12.

Cole, D.R. and Monger, H.C. (1994) Influence of atmospheric CO_2 on the decline of C_4 plants during the last deglaciation. *Nature* 368, 533-536.

Crathorne, B. and Dobbs, A.J. (1990) Chemical pollution of the aquatic environment by priority pollutants and its control. In: Harrison, R.M. (ed.), *Pollution: Causes, Effects, and Control*. 2nd edition. Royal Society of Chemistry, Cambridge, UK, chapter 1.

Crosson, P. (1993) Impacts of climate change on the agriculture and economy of the Missouri, Iowa, Nebraska, and Kansas (MINK) Region. In: Kaiser, H.M. and Drennen, T.E. (eds), *Agricultural Dimensions of Global Climate Change*. St. Lucie Press, Delray Beach, Florida, pp. 117-135.

Dohrenwend, R.E. (1977) *Raindrop Erosion in the Forest. Res. Note No. 24*, Ford Found. Ctr., Mich. Tech. Univ., L'Anse, Michigan.

Dyrness, C.T. (1967) Erodibility and erosion potential of forest watersheds. In: Sopper, W.E. and Lull, H.W. (eds), *Forest Hydrology*. Pergamon Press, New York, pp. 599-610.

Eckholm, E.P. (1976) *Losing Ground*. W.W. Norton and Co., New York.

El-Swaify, S.A. (1987) Soil-based concerns for soil and water conservation research and development in the tropics. In: *Proceedings of the International Workshop on Soil Erosion and its Countermeasures*. Chuan Printing Press Ltd., Part. Bangkok, Thailand, pp. 165-174.

El-Swaify, S.A., Dangler, E.W. and Armstrong, C.L. (1982) *Soil Erosion by Water in the Tropics. Research Extension Series 024*. University of Hawaii Institute of Tropical Agriculture and Human Resources, Hawaii.

FAO, UNEP, UNESCO (1979) *A Provisional Methodology for Soil Degradation Assessment*. FAO, United Nations, Rome.

Fearnside, P.M. (2000) Global warming and tropical land-use change: greenhouse gas emissions from biomass burning, decomposition and soils in forest conversion, shifting cultivation and secondary vegetation. *Climatic Change* 46, 115-158.

Florence, R.G. (1967) Factors that may have a bearing upon the decline of productivity under forest monoculture. *Australian Forestry* 31, 50-71.

Froehlich, H.A. (1988) Causes and effects of soil degradation due to timber harvesting. In: *Degradation of Forested Land: Forest Soils at Risk. Proceedings 10th B.C. Soil Science Workshop: 3-12, Feb. 1986*. Land Mgmt. Rep. 56. Ministry of Forests, British Columbia, Canada.

Garwood, N.C., Janos, D.P. and Brokaw, N. (1979) Earthquake-caused landslides: A major disturbance to tropical forest. *Science* 205, 997-999.

Gornitz, V. and Lebedeff, S. (1987) Global sea-level changes during the past century. *Society of Economic Paleontologist and Mineralogists Spec. Publ. No. 41*, 3-16.

Hagen, L.J. (1976) Windbreak design for optimum wind erosion control. In: *Proceedings from Symposium On Shelterbelts on the Great Plains, Denver, Colorado, April 20-22, 1976*.

Harvey, L.D. (1995) Warm days, hot nights. *Nature* 377, 15-16.

Hashim, G.M. (1988) Soil erosion under tree crops. In: *Proceedings of 5th International Soil Conservation Conference, Bangkok, Thailand*. pp. 28-36.

Hudson, N.W. (1981) *Soil Conservation*. 2nd Edition. Cornell University Press, Ithaca, New York.

Hulme, M. and Viner, D. (1998) A climate change scenario for the tropics. *Climatic Change* 39, 145-176.

IPCC (Intergovernmental Panel on Climate Change) (1990) *Climate Change: The IPCC Assessment*, Houghton, J.T., Jenkins G.J. and Ephraums, J.J. (eds). Cambridge University Press, Cambridge, UK.

Jorgensen, J.R. and Craig, J.R. (1983) Legumes in Forestry: Results of Adaptability Trials in the Southeast. *USDA Forest Service, Southeast Forest Experiment Station Research Paper SE-237*.

Kiehl, J.T. and Briegleb, B.P. (1993) The relative roles of sulfate aerosols and greenhouse gases in climate forcing. *Science* 260, 311-314.

Lal, R. (1987) *Tropical Ecology and Physical Edaphology*. John Wiley, London, 255pp.

Lal, R. (1988) Soil degradation and the future of agriculture in sub-Saharan Africa. *Journal of Soil and Water Conservation* 43, 444-451.

Lal, R. and Greenland, D.J. (eds) (1979) *Soil Physical Properties and Crop Production in the Tropics*. John Wiley, London.

Lau, K.M. and Yang, S. (1996) The Asian monsoon and predictability of the tropical ocean-atmosphere system. *Quarterly Journal of the Royal Meteorological Society* 122, 945-957.

Lawson, T.L., Lal, R. and Oduro-Afriyie, K. (1981) Rainfall redistribution and microclimatic changes over a cleared watershed. In: Lal, R. and Russell, E.W. (eds), *Tropical Agricultural Hydrology*. Wiley, Chichester, UK, pp. 141-151.

Lee, S.W. (1981) Studies on watershed management problems in Taiwan. *Journal of Chinese Soil and Water Conservation* 12, 121-124 (in Chinese).

Lo, K.F.A. (1999) Re-evaluation of rainfall erosivity factor. *Journal of Chinese Soil and Water Conservation* 30, 260-270 (in Chinese).

Lo, K.F.A. and Chiang, S.H. (1996) Impact of land use on environmental quality in Shihmen Reservoir watershed. *Journal of Engineering Environment* 15, 11-18 (in Chinese).

Lo, K.F.A. and El-Swaify, S.A. (1987) Erosion study at Amauulu, Hawaii. *A Progress Report submitted to the Eucalyptus Plantations for Energy Production in Hawaii Project*. BioEnergy Development Corporation, Hawaii.

Loaiciga, H.A., Valdes, J.B., Vogel, R., Garvey, J. and Schwartz, H. (1996) Global warming and the hydrologic cycle. *Journal of Hydrology* 174, 83-127.

Lowdermilk, W.C. (1930) Influence of forest litter on runoff, percolation and erosion. *Journal of Forestry* 28, 474-491.

Luk, S.H. (1979) Effects of soil properties on erosion by wash and splash. *Earth Surface Processes and Landforms* 4, 241-255.

Lundgren, B. (1978) Soil conditions and nutrient cycling under natural and plantation forests in Tanzanian highlands. *Reports in Forest Ecology and Forest Soils No. 31*. Swedish University of Agricultural Science, Uppsala.

Lyles L. (1988) 4. Basic wind erosion processes. *Agriculture, Ecosystems and Environment* 22/23, 91-101.

Lyles, L. and Tatarko, J. (1986) Wind erosion effects on soil texture and organic matter. *Journal of Soil and Water Conservation* 41, 191-193.

Nair, P.K.R. and Balakrishnan, T.K. (1977) Ecoclimate of a coconut + cacao crop combination on the west coast of India. *Agricultural Meteorology* 18, 455-462.

Noguchi, S., Abdul Rahin, N., Zulkifli, Y., Tani, M. and Sammori, T. (1997) Rainfall-runoff responses and roles of soil moisture variations to the response in tropical rain forest, Bukit Tarek, Peninsular Malaysia. *Journal of Forest Research* 2, 125-132.

Patnaik, N. (1978) Soil conservation and watershed management for flood retardation and prevention: A review. *Indian Journal Soil Conservation* 6, 43-55.

Pereira, H.C. (1973) *Land Use and Water Resources*. Cambridge University Press, Cambridge, UK.

Pereira, H.C. McCulloch, J.S.G., Dagg, M. and Kerfoot, O. (1962) Hydrological effects of changes in land use in some East African catchment areas. *East African Agricultural Forestry Journal*, Special Issue No. 27.

Rice, R.M., Rothacher, J.S. and Megahan, W.F. (1972) Erosional consequences of timber harvest: An appraisal. In: *Proceedings from Symposium On Watershed in Transition, Fort Collins, Colorado, June 19-22.* pp. 321-329.

Robinson, C.A., Ghaffarzadeh, M. and Cruse, R.M. (1996) Vegetative filter strip effects on sediment concentration in cropland runoff. *Journal of Soil and Water Conservation* 51, 227-230.

Sedjo, R.A. (1989) Forests to offset the greenhouse effect. *Journal of Forestry* 22, 12-15.

Sidle, R.C., Pearce, A.J. and O'Loughlin, C.L. (1985) *Hillslope Stability and Land Use*. Water Resources Monograph No. 11, American Geophysical Union, Washington, DC.

Sidle, R.C., Tsuboyama, Y., Noguchi, S., Hosoda, I., Fujieda, M. and Shimizu, T. (2000) Stormflow generation in steep forested headwaters: a linked hydrogeomorphic paradigm. *Hydrological Processes* 14, 369-385.

Smit, B. (1987) *Implications of climatic change for agriculture in Ontario*. CCD 87-02, Ottawa, Atmospheric Environment Service, Canada.

Smith, D.D. and Wischmeier, W.H. (1962) Rainfall erosion. *Advances in Agronomy* 14, 109-148, Academic Press, New York.

Spencer, T., Douglas I., Greer, T. and Sinun, W. (1990) Vegetation and fluvial geomorphic processes in Southeast Asian tropical rain forests. In: Thornes, J.B. (ed.), *Vegetation and Erosion*. Wiley, Chichester, UK, pp. 451-469.

Stocking, M. and Elwell, H. (1976) Vegetation and erosion: A review. *Scottish Geographical Magazine* 92, 4-16.

Tejwani, K.G. (1977) Trees reduce floods. *Indian Farming* 26, 57.

Tejwani, K.G. (1979) Soil fertility status, maintenance and conservation for agroforestry systems on wasted lands in India. In: Mongi, H.O. and Huxley, P.A. (eds), *Soil Research in Agroforestry*. ICRAF, Nairobi, Kenya, pp. 141-174.

Tsukamoto, Y. (1975) Effects of forest litter on runoff cycle in a small experimental watershed. In: *Proceedings from The Hydrological Characteristics of River Basin*. IAHS Publication 117, pp. 487-495.

UK Meteorological Office (1998) *Climate Change and Its Impacts*. Web site: www.met-office.gov.uk/sec5/sec5pg1.html.

Vis, M. (1986) Interception, drop size distributions and rainfall kinetic energy in four Colombian forest ecosystems. *Earth Surface Processes and Landforms* 11, 591-603.

Water Resources Planning Commission, Department of Economic Affairs (WRPC, DEA) (1995) A study of water resources impact assessment and appropriate strategies for climatic changes in Taiwan. *Project Report WRPC86-012*, Taipei, Taiwan (in Chinese).

Wiersum, K.F. (1985) Effects of various vegetation layers in an *Acacia auriculiformis* forest plantation on surface erosion in Java, Indonesia. In: El-Swaify, S.A., Moldenhauer, W.C. and Lo, A. (eds), *Soil Erosion and Conservation.* Soil Conservation Society of America, Ankeny, Iowa, pp. 79-89.

Wischmeier, W.H. and Smith, D.D. (1978) Predicting Rainfall Erosion Losses: A Guide to Conservation Planning. *USDA Agricultural Handbook 537*, Washington, D.C.

Wood, H.B. (1977) Hydrological differences between selected forested and agricultural soils in Hawaii. *Soil Science Society of America Journal* 41, 132-136.

Woodruff, N.P. and Siddoway, F.H. (1965) A wind erosion equation. *Soil Science Society of America Procedings* 29, 602-608.

World Bank/FAO (1981) Forestry research needs in developing countries – time for a reappraisal. *Paper for 17th IUFRO Congress, Mimeo 31,* Kyoto, Japan, 6-17 September 1981.

Zobeck, T.M. (1991) Soil properties affecting wind erosion. *Journal of Soil and Water Conservation* 46, 112-118.

Chapter 5

Climate Change Related to Erosion and Desertification: 1. Mediterranean Europe

Giuliano Rodolfi[1] and Camillo Zanchi[2]
[1]*Department of Soil Science and Plant Nutrition, University of Florence, Italy*
[2]*Department of Agronomy and Land Management, University of Florence, Italy*

The combination of recent volcanic eruptions, frequent tectonic adjustments, weak sedimentary deposits and associated soils, and frequent high-intensity storms make the Mediterranean basin vulnerable to surface erosion by water and related climate changes. The region has experienced a long history of human disturbances that have exacerbated soil erosion, beginning with forest removal in ancient times up to the most dramatic impacts arising from socio-economic transformations and agricultural mechanization following World War II. With the predicted temperature increases, there will probably be a stronger and more prolonged drought period and a northward extension of arid conditions (desertification). Soil structure may be degraded due to reductions in organic matter and increased salinization. The resultant decrease in aggregate stability would lead to reduced water retention and increased erodibility of soils, especially when cultivated. Although the impact of climate change on both natural vegetation and crops would probably be moderate, it may become significant in marginal areas due to increased aridity and the northward shifting of vegetation belts. However, it is the interaction of changing agriculture practices in the region with potential climatic warming together with their influence on other land uses and socio-economic conditions that will determine the trajectory of land degradation. Abandonment of marginal cultivated lands without maintaining conservative practices could lead to rapid deterioration, massive erosion, and irreversible conditions for future land use. Pressures from tourism in the region will increase, exerting impacts on land use allocation, water availability, and tourist structures and infrastructures. Such anthropogenic effects will certainly dominate over climate change effects in southeast Mediterranean Europe, due to a continuing regional demographic shift.

Introduction

The Mediterranean basin is considered to be a unique environment. Although centrally located in the so-called middle latitudes, it is not particularly representative of this region. In the other continents of the mid-latitudes, landmasses extend continuously; however, the Mediterranean basin ("sea among lands" in Latin) lies between Europe and Africa, which thus influences the climate of the coastal regions. The Mediterranean basin is protected by high mountainous barriers and is largely protected from cold winds from the north or northeast; therefore, temperature fluctuates less widely than in larger land masses at the same latitude in other continents. Another peculiarity of the Mediterranean basin is the significant relief differences encountered even in proximity to the coastal belt. The lack of extensive plains and the significance of mountain ridges in the basin contribute to the complexity of the climate. The influence of human activity for a period of more than 4000 years has added to the complexity and variability of the present environmental and landscape conditions in the Mediterranean basin and has been a strong force in modifying the landscape.

Of the various Mediterranean nations, Greece, Italy and the former Yugoslavia occupy and control two-thirds of the entire Mediterranean shoreline (Milliman, 1992a). Peninsular Italy is the most representative of these shorelines as it diverges sharply from continental Europe and extends southward more than 10° in latitude (from about 45°N to about 34°N). Therefore, most of our discussions related to the Mediterranean basin will focus on peninsular Italy or the big islands (Sicily and Sardinia) that are located in the middle of the Mediterranean Sea.

A complex geological history

From the point of view of endogenous geophysical processes, the Mediterranean basin is one of the most active areas of the Earth. Located between the northern margin of the African plate and at the southern margin of the European plate, it is prone to frequent tectonic adjustments that represent the late phase of the alpine orogenesis. Frequent, sometimes catastrophic, earthquakes have occurred throughout the region, save only the more ancient and stable areas such as Sardinia and Apulia. A series of recently extinct volcanoes are aligned parallel to main tectonic structures of the Apennine, while volcanoes rise along the coast (e.g., Vesuve, Etna) or in the open sea (e.g., Stromboli, Vulcano).

Sedimentary bedrock dominates the region with limited outcrops of basement crystalline rocks and more recent intrusive batholiths. Along the coasts, the postorogenic (plio-quaternary) cycles of transgression and regression have allowed the sedimentation of very thick series (more than 1000 m) of clay, sand and conglomerates. These deposits are generally poorly cemented and prone to degradation. The deposits and associated soils form the typical 'badland' landscapes (Italian: *calanchi*) throughout wide areas of the Mediterranean basin (Fig. 5.1). Throughout the inner continent, outcrops of weaker formations are common, either due to their intrinsic composition (e.g., shales, marls) or to intense tectonic events (e.g., faulting, overthrusting) or both. In these areas, the effects of infiltrating water during rainstorms on the initiation of large mass movements must be considered together with the erosion due to surface wash.

Fig. 5.1. A characteristic erosional ('badlands-like') landscape, very common on the Apennine piedmont in Italy: the "calanchi" (photo: G. Rodolfi).

An unusual climate

Since the Mediterranean climate is located between the temperate and tropical zones, it exhibits characteristics of both: winters are typical of the temperate zone and summers are more tropical (Pinna, 1977). Regional climate is determined by characteristics of the large-scale atmospheric circulation, interactions between airflows, orography and sea-land contrasts, as well as by more local effects (Wigley, 1992). The regional climate is classified as *Csa* type (subtropical with dry summers) according to the Köppen system. Rainfall is usually sparse and concentrated during the coldest months. The Mediterranean regime of transition, north of the 41° parallel, has two distinct rainfall peaks (late autumn and spring); the typical Mediterranean regime has a longer dry period with rainfall mainly occurring during the three winter months: at Palermo (Sicily) about 50% of the total rain falls from December to February. Individual rainfall events are generally characterized by short duration and high intensity. In Versilia (seaward Tuscany) catastrophic flash flooding resulted from a prolonged (477.4 mm of rain occurred in 15 hours) and intense (peak intensities of 158 mm h^{-1} and 30.8 mm in 5 minutes) storm on June 19, 1996 (Rapetti and Rapetti, 1996). On November 13-15 1999, a rainfall event of 330 mm in 12 hours in the town of Assemini, about 20 km west of Cagliari (southern Sardinia), caused a strong flash flood as well as mudflows and debris flows. During the same period, 214 mm of rain fell during a 4-hour period in Sarrabus (Sardinia) 20 km to the east of Cagliari. These recent events illustrate the significance of the geomorphic influence of rainfall in the region south of the 41° parallel.

Fig. 5.2. Concentrated overland flow during short, but extremely intense rains, deeply eroded a recently cleared hillslope of a *Pinus halepensis* forest in Mediterranean Spain (photo: G. Rodolfi).

The temperature regime in the region indicates a strong contrast between winter and summer. Winter is generally mild. The average temperature in January fluctuates between 6°C in the northern sector to 12°C in the southern sector, both sectors experiencing frequent perturbations due to cold air from the north. Summers are very hot, dry, and characterized by a prolonged insolation in southern Italy (Benincasa *et al.*, 1991). The average temperature of the hottest month exceeds 22°C, with daily maximums exceeding 40°C. Average annual temperatures along the coast are below 15°C, but these increase considerably in the interior portion of the basin. The vegetative growing season spans 240 to 300 days (Perrin, 1954). In the southern and eastern sectors of the basin the climate becomes more arid-like and is more typified by the dry steppe zone of the *Köppen* classification system.

At the global scale these climatic conditions are an anomaly because in other continental regions rain typically occurs during summer (exceptions being deserts, where little rain falls and coastal areas, where rain occurs every month) (Pinna, 1977). The concentration of short-duration, high-intensity rainfall events in the region strongly influences geomorphic processes. As a consequence, rainfall in the Mediterranean environment imparts a very high level of rain erosivity to the landscape (Fig. 5.2).

Because of its unique location, the Mediterranean basin, more than adjacent regions, has witnessed the effects of the climatic fluctuations that have occurred since the early Quaternary. Climatic variations in the region have affected the relief and the composition and distribution of vegetation. These climatic shifts have been both long-term (i.e., alternating Pleistocene glaciations) and short-term (several-year cycles) and have impacted human evolution in the region as well as, more recently, the extent and nature of human activities, including land use.

Ancient human influences

The Mediterranean basin is considered the cradle of the human civilization. Even before the birth of the most ancient civilizations (as Sumer and Akkad), primitive agriculture was modifying the natural landscape. Early forest removal exposed the soil for prolonged periods, exacerbating erosive processes. The widespread outcrops of bare calcareous bedrock in southern Italy and in Greece (Fig. 5.3) are examples of this prolonged degradation. In spite of this early degradation, the precarious balance between human activities and natural processes was largely maintained except for the consequences of some episodic rainfall events. However, due to socio-economic transformations that occurred after the Second World War, including both rapid industrial and urban development as well as consequent changes in the agricultural systems (e.g., crop intensification and mechanization on a vast scale), land degradation increased.

Moreover, increasing use of the Mediterranean coastline by tourists has severely impacted the region. In 1984, 100 million tourists visited important resort areas concentrated on 200 ha of land and consumed 569 m^3 of fresh water (UNEP, 1989). Since in the year 2000, 120 to 180 million people used these same Mediterranean beaches, this number is expected to increase up to 170–340 million by 2025 (UNEP, 1987).

Fig. 5.3. A "calanchi" landform showing the rate of vertical erosion in a Pliocene silty-clayey formation bordering peninsular Italy. The brick-kiln in upper foreground, constructed by Romans on a wider slope, is now standing on a sharp divide (photo: G. Rodolfi).

Present-day desertification and its causes

UNEP (1991, 1992) describes desertification is "...the degradation of land in arid, semi-arid and dry sub-humid areas resulting from various causes and mainly to climatic variations and adverse human impact...". This definition includes climatic, pedo-geomorphic and physical conditions, as well as human activity, which are influenced by both individual and cumulative socio-economic factors, including political decisions.

It is well known that the degradation and desertification in Mediterranean Europe are processes attributed to human activities and exacerbated by climate. The importance of these phenomena is clearly illustrated in southern Europe where desertification threatens more than 60% of the landscape including the central and southeastern regions of Spain, southern Portugal, central and southern Italy (including Sicily and Sardinia), and large areas of Greece and Crete (UNEP, 1991; Middleton and Thomas, 1997).

As noted earlier, during the past 40 years fragile ecosystems in the Mediterranean basin have been subjected to a rapid and progressive intensification of agricultural practices, promoted either by the evolution and import of mechanization or by the quest for greater agricultural income (Guidoboni, 1998). This agricultural expansion has resulted in larger fields with little attention to conservative practices and over-cultivation and intensification of crop rotation (Fig. 5.4). Additionally, more monoculture crops are being grown with a correspondent reduction of pasturelands and related inputs of organic matter into the soil. As an example, between 1965 and 1984, pastureland in Mediterranean Europe declined by some 3 million ha, representing about 7% of the total cultivated area (Perez-Trejo, 1994). Contemporarily, there has been a progressive loss of the most fertile agricultural land due to increased urbanization, establishment of recreational areas and infrastructures, and relocating agricultural practices in more fragile hillslope areas (Rodolfi, 1988). The most significant land use changes that have affected hillslope agricultural lands in the region are:

- widespread use of fertilizers and agro-chemicals in general;
- abandonment of old conservation practices (such as bench terracing, contour ditching, hill-riding (Italian: "cavalcapoggio");
- increase of irrigated land area without application of specific conservation measures (i.e. *tile* drainage), particularly where the irrigation water used is of fair or poor quality;
- progressive reduction of water availability and quality that results from both competition with non-agricultural uses and increased pollution (e.g., increasing urban and industrial wastes, and salt concentrations);
- progressive increase of fires, both in number and frequency.

Fig. 5.4. Severe rill erosion in silty-clayey soil on a wide, recently ploughed, slope in Basilicata (southern Italy) (photo: G. Rodolfi).

Although all natural waters are of hydrological interest, from the perspective of consumptive use, only those waters available at the required site and time, as well as in adequate quantity and quality, are useable. In Table 5.1 selected hydrological parameters for the major coastal areas of the Mediterranean are reported. In the context of these underlying hydrologic conditions, there has been and still appears to be a general increase in water and wind erosion processes, soil and water pollution, and soil salinization in the region.

Table 5.1. Hydrological parameters of the main coastal areas of the Mediterranean, except Greece (data from Henry, 1977; Lindh, 1992).

Coastal locality	Precipitation (mm $year^{-1}$)	Aridity (months $year^{-1}$)	Runoff (mm $year^{-1}$)	Runoff coefficient
Eastern Spain	200 – 400	3 – 7	100	0.1
Western Italy and Liguria	800 – 1200	2 – 3	400 – 500	0.5
Former Yugoslavia	> 1500	1 – 2	300	0.6
Southern Mediterranean	100 – 400	4 – 5	100 (Tunis)	0.1

Because of land use changes in the region, including human misuse/exploitation of natural resources (particularly soil and water), as well as climatic fluctuations and modifications, it is evident that the environmental sensitivity of Mediterranean Europe is particularly prone to severe erosion and degradation. Consequently, the resulting desertification is a complex multi-causal phenomenon in such dynamic systems (Ferrari and Zanchi, 1979). Much focus has recently been directed on identifying indicators of desertification, not only to forecast its early stages, but also to provide mechanisms to combat desertification and to reduce its consequences.

Global Warming: Natural or Human-induced Event?

It is clear that the Mediterranean basin is one of the most complex environments on Earth, due to its geologic, climatic, and physiographic conditions, as well as historical and contemporary human influences. All these linked conditions have been subjected through time to fluctuations of different amplitudes. For example, since the 19th century the mean annual temperature of the world has increased about 0.5°C. However, temperature trends have not been continuous either in time or space, thus the greenhouse effect may have been masked by natural climatic variability (Wigley, 1992). In the past 20 years the western Mediterranean has become warmer while the eastern Mediterranean has become cooler. Thus, it is impossible to currently ascertain whether the Mediterranean basin as a whole is following global trends. Similarly, it is currently difficult to state if the present-day climate changes in the Mediterranean are the consequences of human activity or the effect of natural processes, like those that occurred in the recent past (Fig. 5.5).

Fig. 5.5. A landscape in southwestern Spain that is almost completely desertified. It is difficult to ascertain whether this increasing aridity and strong erosion is due to natural climate oscillations, or a consequence of exploitation of this fragile ecosystem (photo: G. Rodolfi).

Currently, water and wind erosion are considered among the most important causes of soil degradation and, consequently, of land desertification. Even though soil erosion is a natural process, human activity and the evolution of agricultural systems have recently accelerated this process. It is well known that increases in erosion cause a progressive degradation of soil fertility and productivity due to the reduction of soil depth and the removal of organic matter. Moreover, the degradation of soil structure and compaction of surface layers causes decreases in infiltration capacity and water retention as well as potentially irreversible effects on the potential for vegetation recovery. In order to maintain soil fertility and productivity, there must be a balance between human activity and the environment. This balance can be identified by the relationship between the amount of eroded soil and the amount of soil regenerated by pedogenetic processes (Perez-Trejo, 1994). These values are extremely variable and difficult to quantify because they depend on specific inter-relations among climate, substrata, vegetation, and agronomic techniques. The estimated time to regenerate 1 cm of soil in the Mediterranean region is about 12 years, for the most favourable conditions and 120-400 years, in certain natural environments (Giordani and Zanchi, 1995). In addition to the on-site effects of surface erosion (e.g., decrease in land productivity, damage to infrastructures), the eroded sediment represents a non-point source of pollution (e.g., Tim and Jolly, 1994).

Consequences of Global Warming

Besides the anthropogenic factors already considered, this paper focuses on how various land degradation processes may be influenced by the global warming trends articulated by the Hadley Centre Meteorological Office (1998). For the Mediterranean region, results from general circulation models reflect global warming trends although more meteorological observations are needed to draw statistically reliable conclusions. Changes in precipitation, however, are difficult to predict due to the spatial variability of rainfall and the lack of data for oceanic areas (Wigley, 1992). Increasing sea surface temperature of both the western Mediterranean and northern Atlantic could increase atmospheric humidity thereby causing more rain.

In the next 50 years, the mean annual temperature of the Mediterranean region is forecasted to increase by about 3°C (Hadley Centre, 1998). Evapotranspiration is consequently predicted to increase to 620 mm year^{-1} (about a 6% increase), although unevenly distributed (Le Houerou, 1992). Expected changes in precipitation, evapotranspiration and runoff are reported in Table 5.2. Periods of extreme temperatures, causing forest fires and reduced availability of water for vegetation are expected (Wigley, 1992; Imeson and Emmer, 1992).

Table 5.2. The absolute and relative changes in annual precipitation, evapotranspiration and runoff (the values refer to data within latitudes 35 to 45°) (source: Oljenik, 1988; Lindh, 1992).

	Precipitation (mm)	Evapotranspiration (mm)	Runoff (mm)
Present	726	583	141
Future	807	620	187
Change	81	37	46
Increase in %	+ 11	+ 6	+ 33

Consequences on geomorphic processes

Based on the previous discussion we can deduce that many geomorphic hazards, leading to desertification, are active in Mediterranean Europe. Ignoring the endogenous geophysical processes related to crustal dynamics, careful attention must be paid to exogenous (or climate-dependent) processes, either as triggers of geomorphic hazards or as regulators of their evolution. Geomorphic hazards can be defined as "...those events or processes, natural or man-induced, that cause change in earth-surface characteristics detrimental to Man and his activities. They form a sub-set of the broader range of natural hazards that represent a world problem of growing importance, in which the cost to mankind is measured in billions of dollars annually..." (Embleton et al., 1989). A classification of the geomorphic hazards in which climatic factors play a dominant role is represented in Table 5.3.

Table 5.3. A classification of the main geomorphological hazards, due to exogenous processes (adapted from Embleton *et al.*, 1989).

Major process	Potentially hazardous Earth surface processes	Possibly man-induced	Possibly sudden & catastrophic	Ability of man to control	Predictability
Gravity	Mass movements	Yes	Yes	Moderate	Good, Moderate
Fluvial	Flooding Bank erosion	Yes	Yes	Good, Moderate	Good, Moderate
Soil erosion by water	Splash, interrill, rill, gully	Yes	Yes	Good, Moderate	Good, Moderate
Glacial	Advancing glacial Retreating glacial	No ?	Yes No	Poor	Good, Moderate
Nival	Avalanches	Yes	Yes	Moderate	Good, Moderate
Marine	Coastal erosion Silting, accretion Salt-water intrusion	Yes	No	Good, Moderate	Good
Soil erosion by wind	Dune migration Sand-silt covers	Yes	No	Moderate, Poor	Good

It is apparent that most of these geomorphic hazards may be human-induced, somewhat controllable (unless catastrophic) and moderately predictable (Table

5.3). Once again, the role of human activity as geomorphic agent appears to be of paramount importance in regulating the entire denudation-aggradation cycle (Nir, 1983). It is also clear how a climatic change can influence all of these processes and hazards and their linkages.

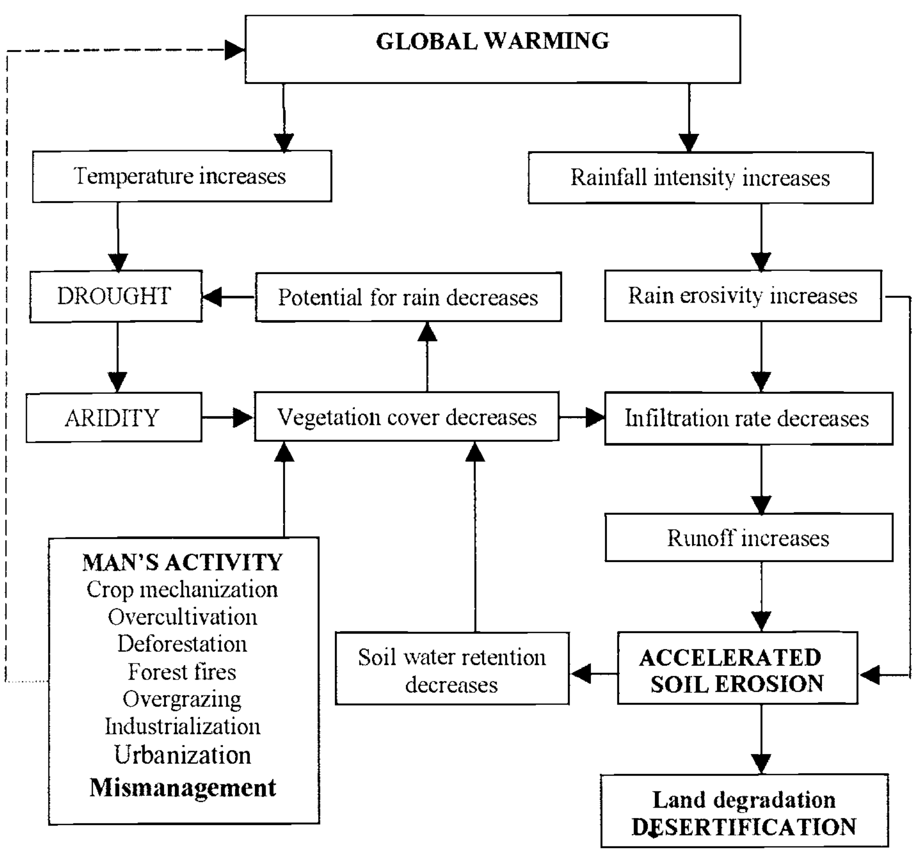
The climate of Mediterranean Europe is characterized by high rainfall intensities with a very irregular distribution throughout the year, prolonged periods of drought related to periods of high temperature, high evapotranspiration rates, and strong winds. Moreover, the arid conditions, mainly during late spring, summer and early autumn, as well as the reduced drainage, lead to an accumulation of salt at the soil surface. Such saline conditions strongly influence the dominant plant communities (*macchia mediterranea*) as well as farming systems and affect the ability of these systems to protect against erosive processes (Fierotti and Zanchi, 1998) (Fig. 5.6).

Fig. 5.6. Sardinia Island (Italy) – degraded Mediterranean bush ("*macchia*") and shallow soils are witness of a very precarious balance between environmental parameters (photo: G. Rodolfi).

If we accept the predicted temperature-increase scenarios, there will likely be a stronger and more prolonged period of drought and a northward extension of arid conditions into areas previously not subjected to such extended dry periods (Perez-Trejo, 1994) (Table 5.4). These conditions may lead to degradation of plant communities or, at least, to reduced growth and capacity of sustaining cultivation. Such a situation may trigger a series of ecosystem feedbacks that can be summarized as follows:

- interactions between reduced vegetation cover and consequent reduced potential for rainfall;
- reduced effect on soil protection against erosion;
- reduced infiltration, more runoff, and less water retention and availability for vegetation.

Table 5.4. General scheme showing how global warming (natural and/or human-induced) could lead to desertification in Mediterranean Europe.



In many cases these feedbacks lead to increased grazing (Fig. 5.7), with associated soil compaction and soil erosion, thus accelerating the degradation of both soil and vegetation. Consequently, the composition of plant communities is subject to change, possibly leading to a reduction of those species that are useful for maintaining soil structure and fertility. Increased aridity also necessitates more water for irrigation, which generates conflicts due to the general reduction in total water availability and with changes in water allocation (e.g., competition for water from industry, tourism, and urbanization or because of increases in the price of water).

Fig. 5.7. Sheep-tracks are the most evident signs of overgrazing on the slopes, leading to complete land desertification in Murcia, Spain (photo: G. Rodolfi).

The increased use of irrigation water due to poor management (particularly because of the lack of adequate drainage) could lead to problems, such as waterlogging, salinization and alkalinization, in many areas. These problems could also be exacerbated by decreased quality of the irrigation water. Irrigation agriculture, particularly the related costs for establishment and maintenance, leads to compulsory changes in cropping patterns, favouring the high-value cash crops. Such systems are not necessarily compatible with soil and water conservation.

Longer periods of aridity are typically associated with shorter rainy periods with higher intensity, thus leading to higher rates of erosivity and greater erosion potential (Imeson and Emmer, 1992), increased surface runoff and flooding hazard, and siltation in streams, rivers and reservoirs. It is well known that increases in rainfall intensity and associated increases in the raindrop energy provide the primary energy source for erosion. A study in central Italy observed that rainfall energy increased significantly with increases in temperature (Zanchi and Torri, 1980; Zanchi, 1988a,b). Rain splash modifies soil physical properties, including breaking soil aggregates, compacting the soil surface, forming a surface crust, clogging surface soil pores, and, thus, reducing infiltration. As a consequence, soil particles are more easily detached and transported downslope by the increased runoff volume, thus producing a higher potential for soil erosion and strongly modifying the hydrological cycle within the soil. The likely climatic trends point to a more frequent occurrence of extreme events that ultimately could generate the more serious erosion and consequent onset of desertification (e.g., 39°C air temperature in southern Italy in October 1999; tropical-like rains in Sardinia and southern France in November 1999). Numerous researchers (e.g., MEDALUS, 1993; Perez-Trejo, 1994; Fierotti and Zanchi, 1998; Rubio and Bochet, 1998) observed that more than 70% of the total annual soil loss comes from a few extreme and unpredictable rainfall events that cannot properly be monitored during typical short-term studies or measurements (Poesen *et al.*, 1996).

Consequences on soil - vegetation - land use complex

Certain aspects of the strong interrelationships among climate, soil, and vegetation (either natural or cultivated) and erosion, have been studied. Concerning soil, the main effect of climate change will be on both the organic matter and chemical composition, which could lead to an irreversible degradation of soil structure. In cultivated soils this translates to the production of smaller and less stable aggregates, thus decreasing water retention and resistance of soil particles to detachment. Therefore, the erodibility will generally increase, especially for the most susceptible soils (e.g., silty soils and soils prone to dispersion) (Chisci and Zanchi, 1980; Imeson and Emmer, 1992). The impact of climate change on both natural vegetation and crops could be moderate; nevertheless, it may become significant in areas which are marginal for soil or vegetation conditions due to increased aridity and the northward shifting of vegetation belts as well as cool-sensitive crops (olives, vegetables and citruses) (Le Houerou, 1992).

Moreover, changes in agriculture as a result of climatic warming will also affect other land uses and socio-economic conditions. Abandonment of marginal cultivated lands without maintaining conservative practices could lead to rapid deterioration, massive erosion, and irreversible conditions for future agriculture.

Pressures from tourism in Mediterranean Europe, attributed to the wide variety of scenic landscapes, ancient remains, and the favourable climate will probably increase, exerting greater environmental impacts, particularly on land use patterns, water availability, and finally on the ever-growing tourist structures and infrastructures. Along the northern coast of Palermo (Sicily) at least 60% of the land has been partially or totally covered by concrete, and between 80% and 90% of the best or most fertile soils have been destroyed and then desertified (Raimondi *et al.*, 1995). Similar situations have occurred in several other Mediterranean areas (Aru *et al.*, 1983).

Table 5.5. Utilization of surface water and groundwater in northern countries of the Mediterranean area in percent and total available water resources in m^3 per inhabitant (adapted from Ennabli, 1982)

Country	Per cent withdrawal		Total available resource (m^3 per inhabitant per yr)
	Surface water	Ground water	
France	50	50	3400
Italy	87	13	2990
Spain	70	30	2900

Another consequence of the increased tourist trade is the reduction in water availability due to the increases in water consumption. Table 5.5 illustrates the available water sources and demands in the three most populated countries of Mediterranean Europe. Due to excessive pumping of groundwater, regional water tables have been lowered and salt water has intruded into aquifers in coastal regions. These groundwater impacts may eventually lead to the abandonment of the coastal cultivated land.

Consequences of sea-level rise

During the past 100 years there has been a general rise in sea level between 10 and 20 cm. In more recent years, increased atmospheric CO_2 levels (≈ 30%) and mean global temperatures (≈ 0.5°C) have not resulted in similar increases in sea level (Milliman, 1992b). One of the most worrying prospects of global warming is the probable sea-level rise, no later than 2025–2030, when the CO_2 content in the atmosphere is expected to double. This sea-level rise due to both thermal expansion of the oceans and melting of polar ice caps (e.g., Antarctica and Greenland), is forecasted to be 12–18 cm; this range could increase up to 25–40 cm in coastal areas subject to subsidence (e.g., the Nile and Po deltas; Fig. 5.8) due to water, oil or gas pumping (Milliman. 1992b). According to Hadley Centre Meteorological Office (1998) the global mean sea-level rise by 2050 is predicted to reach 21 cm. This implies that all coastal plains of the Mediterranean belt will undergo a severe risk of flooding (Imeson and Emmer, 1992) or at least of waterlogging related to regional increases in the groundwater tables. Another consequence will be the growing intrusion of saline waters into the coastal plains (Lindh, 1992); such intrusion will rapidly decrease soil productivity. Moreover, the prolonged stagnation of water will promote certain diseases, such as malaria. Shoreline retreat, already affecting large stretches of Mediterranean beaches, will expand. Therefore, coastal settlements, tourist facilities and industrial sites may have to be abandoned. Relocation of such developments in the interior will increase the impermeable surfaces already associated with existing urban development and will push agriculture into neighbouring hillslopes that are more susceptible to soil erosion. The increasing impermeable land area (e.g., roofs, asphalt surfaces, roads) will progressively reduce drainage-basin response time: the time between the maximum rainfall intensity and peak discharge for a storm. This increased energy of flowing water will lead to downcutting and lateral erosion in the headwaters; in the lower catchment, larger amounts of bedload sediment will be deposited. Consequently, the peak discharge of the rivers, as well as the risk of flooding, will increase. Thus, a new cycle of land degradation may be triggered.

Fig. 5.8. The Po River Delta (Italy): a former cultivated area now invaded by lagoonal waters, following a widespread land subsidence due to exploitation of deep water and gas (photo: G. Rodolfi).

Conclusions

Based on this review it is evident that a general deterioration of the present-day environmental conditions in Mediterranean Europe is to be expected within the next 50 years. Additionally, we must take into account all the increasing consequences of anthropogenic pressures. In the southern and eastern sectors of the Mediterranean basin, population has doubled every 25 years since 1950; in 2050 it is forecast to reach about 1.5 billion inhabitants. This pattern is in sharp contrast with the northwestern sector, where an exodus of the resident population has already begun and surely will increase. Therefore, the consequences of global warming will be less evident for most parts of southeastern Mediterranean Europe, in contrast to the socio-economic changes triggered by this expected demographic shift (Le Houerou, 1992). Humans will become the main geomorphological agent, more and more responsible for environmental degradation.

References

Aru, A., Baldaccini, P., Malquori, P., Melis, R.T. and Vacca, S. (1983) *Il consumo delle terre a causa della espansione urbana del territorio intorno a Cagliari*. Institute of Geology, Univeristy of Cagliari.

Benincasa, F., Maracchi, G. and Rossi, P. (1991) *Agrometeorologia*. Ed. Patròn, Bologna.

Chisci, G.C. and Zanchi, C. (1980) The influence of different tillage systems and different crops on soil losses on hilly silty-clayey soils. In: Morgan, R.P.C. (ed.), *Soil Conservation: problems and prospectives*. John Wiley & Sons, New York.

Embleton, C., Federici, P.R. and Rodolfi, G. (1989) Geomorphological hazards: characteristics and human response. *Supplementi di Geografia Fisica e Dinamica Quaternaria* 2, 1-3.

Ennabli, M. (1982) *Etude sur les ressources en eau dans le bassin Méditerranéen*. Plan Bleu.

Ferrari, G.A.F. and Zanchi, C. (1979) La desertificazione, un fenomeno sempre più intenso. *L'Italia Agricola* 116, 25-35.

Fierotti, G. and Zanchi, C. (1998) Agricultural practices and soil fertility degradation. *Int. Seminar on Indicators for Assessing Desertification in the Mediterranean*, Porto Torres, September 1998.

Giordani, C. and Zanchi, C. (1995) *Elementi di Conservazione del Suolo*. Ed. Patròn, Bologna.

Guidoboni, E. (1998) Ricostruzione storica dei fenomeni di desertificazione e di degrado del territorio in Italia. *ENEA, Report SGA/RPT 197/98*, Bologna.

Hadley Centre Meteorological Office (1998) *Climate Change and its Impacts*. Report, Crown Copyright.

Henry, P.M. (1977) La Méditerranée: un microcosme menacé. *Ambio* 6, 135-148.

Imeson, A. and Emmer, I.M. (1992) Implication of climatic change on land degradation in the Mediterranean. In: Jeftic, L., Milliman, J.D. and Sestini, G. (eds), *Climatic Change and the Mediterranean*. Edward Arnold, London, pp. 95-128.

Le Houerou (1992) Vegetation and Land Use in the Mediterranean basin by the year 2050: a prospective study. In: Jeftic, L., Milliman, J.D. and Sestini, G. (eds), *Climatic Change and the Mediterranean*. Edward Arnold, London, pp. 175-232.

Lindh, G. (1992) Hydrological and water resources impact of climate change. In: Jeftic, L., Milliman, J.D. and Sestini, G. (eds), *Climatic Change and the Mediterranean*. Edward Arnold, London, pp. 58-93.

MEDALUS (1993) *Mediterranean Desertification and Land Use: Executive Summary*. Medalus, King's College, London.

Middleton, N. and Thomas, D. (eds) (1997) *World Atlas of Desertification*. 2nd edition. UNEP, Arnold Publication, London.

Milliman, J.D. (1992a) An overview. In: Jeftic, L., Milliman, J.D. and Sestini, G. (eds), *Climatic Change and the Mediterranean*. Edward Arnold, London, pp. 1-14.

Milliman, J.D. (1992b) Sea level response to climate change and tectonics in the Mediterannean Sea. In: Jeftic, L., Milliman, J.D. and Sestini, G. (eds), *Climatic Change and the Mediterranean*. Edward Arnold, London, pp. 45-57.

Nir, D. (1983) *Man, a geomorphological agent. An introduction to Anthropic Geomorphology*. Keter Publ. House.

Oljenik, J. (1988) Present and future estimates of evapotranspiration and runoff for Europe. *Working Paper WP/88/037*, IIASA, Luxembourg.

Perez-Trejo, F. (1994) Desertification and Land Degradation in the European Mediterranean. *European Union Report 14850*, VI, Bruxelles.

Perrin, H. (1954) Sylviculture (Tome II) : Le Traitement des Forêts, Théorie et Pratique des Techniques Sylvicoles. Italian translation by Bernetti, G. (1986) *Selvicoltura. Acc. It. Scienze Forestali*.

Pinna, M. (1977) *Climatologia*. Unione Tipografica Editrice Torinese, Torino.

Poesen, J.W., Boardman, J., Wilcox, B. and Valentin, C. (1996) Water erosion monitoring and experimentation for global change study. *Journal of Soil and Water Conservation* 2, 386-390.

Raimondi, S., Dazzi, C. and Fierotti, G. (1995) *Consumo del suolo e modificazioni indotte dall'urbanizzazione nella fascia costiera occidentale di Palermo*. Atti Conv. SISS, Cagliari.

Rapetti, C. and Rapetti, F. (1996) L'evento pluviometrico eccezionale del 19 Giugno 1996 in Alta Versilia (Toscana) nel quadro delle precipitazioni delle Alpi Apuane. *Atti Società Toscana Scienze Naturali, Memorie, Serie A* 103, 143-159.

Rodolfi, G. (1988) Geomorphological mapping applied to land evaluation and soil conservation in agricultural planning: some examples from Tuscany (Italy). *Z. Geomorph. N.F., Suppl. Bd.* 68, 155-174.

Rubio, J.L. and Bochet, E. (1998) Indicators for Desertification Assessment. *Int. Seminar on Indicators for Assessing Desertification in the Mediterranean*. Porto Torres, Sept. 1998.

Tim, U.S. and Jolly, R. (1994) Evaluating agricultural nonpoint-source pollution using integrated geographic information systems and hydrologic/water quality model. *Journal of Environmental Quality* 23, 25-35.

UNEP (United Nations Environment Programme) (1987) *Preliminary report on Blue Plan scenarios. (UNEP/WG. 171/3)*. Blue Plan, Sophia Antipolis.

UNEP (United Nations Environment Programme) (1989) State of the Mediterranean marine environment. Jeftic, L. (ed.), *MAP Techn, Rep. Series*, 28, Athens.

UNEP (United Nations Environment Programme) (1991) *Status of desertification and Implementation of United Nations Plan of Action to combat Desertification*. UNEP, Nairobi.

UNEP (United Nations Environment Programme) (1992) *World Atlas of Desertification*. 1st edition. Edward Arnold Publication, London.

Wigley, T.M.L. (1992) Future climate of the Mediterranean basin with particular emphasis on changes in precipitation. In: Jeftic, L., Milliman, J.D. and Sestini, G. (eds), *Climatic Change and the Mediterranean*. Edward Arnold, London, pp. 15-44.

Zanchi, C. (1988a) The cropping pattern and its role in determining erosion risk: experimental plot results from the Mugello valley (Central Italy). In: *Proceedings IAHS Symp., Porto Alegre (Brasile)*. IAHS Publication No. 174, 139-145.

Zanchi, C. (1988b) Soil loss and seasonal variation of erodibility in two soils with different texture in Mugello valley (Central Italy). *Catena Suppliment* 12, 167-173.

Zanchi, C. and Torri, D. (1980) Evaluation of rainfall energy in Central Italy. In: de Boodt, M. and Gabriels, D. (eds), *Assessment of Erosion*. John Wiley & Sons, New York, pp. 133-142.

Chapter 6

Climate Change Related to Erosion and Desertification: 2. Africa

Heinz R. Beckedahl
Faculty of Science and Agriculture, Department of Geography, University of Natal, Pietermaritzburg, South Africa

As much as 87% of the African continent, north of the equator, is affected by accelerated soil erosion. In these developing regions, soil erosion and the associated environmental consequences directly affect the livelihood of the inhabitants, frequently resulting in starvation and even death of local inhabitants. However, soil erosion rates are highly variable, from as low as 1.5 t^{-1} ha^{-1} $year^{-1}$ to an excess of 250 t^{-1} ha^{-1} $year^{-1}$. It is unlikely that this large variance will change in the near future as a result of climate change. What will change is the rate of soil loss for specific regions and districts. Erosion rates in Africa in the future are likely to reflect demographic factors even more strongly than before. The nature and extent of changes in erosion rates will reflect the awareness of people living in these areas to the inherent risks and whether this impacted population will have the human and material resources to implement effective soil conservation measures. Thus the role of humans in environmental degradation appears to be twofold: (i) their role as principal factors or contributors of global change; and (ii) land misuse caused by humans. Together these factors may contribute to a progressive downward spiral of environmental degradation.

Background

Recent estimations indicate that as much as 87% of Africa north of the equator, including the entire Mediterranean zone of North Africa, is affected by accelerated erosion, a global process which is responsible for the removal of some 75 billion metric tonnes annually (FAO, 1983; Millington, 1992; Pimentel *et al.*, 1995; Middleton and Thomas, 1997). The consequences of this degradation are illustrated by the projected decline in the global per capita arable land area from 0.3 ha in 1986, to 0.23 ha by 2000 and eventually to 0.15 ha by 2050 (Brown *et al.*, 1990). Sub-Saharan Africa appears to be experiencing similar magnitudes of decline in arable land per capita from 5.5 ha (in southern Africa) in 1970 to approximately 1.5 ha by 2000 (Beckedahl, 1998) with annual soil loss for South Africa in the early 1980s varying between 300 and 400 t per annum, or some 10 t per capita per annum (Beckedahl *et al.*, 1988). Such estimates are gross generalizations and hence of limited value, especially considering the large range of present day erosion rates cited in the literature. For South Africa alone, these rates vary from as low as 0.5 t^{-1} ha^{-1} $year^{-1}$ to more than 110 t^{-1} ha^{-1} $year^{-1}$ (Boucher and Weaver, 1991). Similarly large ranges occur in many other African nations (Stocking, 1984), thus it is of limited benefit in such a review to cite specific rates of soil loss. While the specific figures for Africa can be obtained from the cited references, they suggest that the mean rate of soil loss in Africa may (at least regionally) exceed the generally accepted rate of soil formation of 10 t^{-1} ha^{-1} $year^{-1}$ by a factor of more than ten (see Bork, 1988).

To emphasize the major differences between soil erosion in Africa and the developed nations of Europe and North America, soil loss data related to land degradation in Africa must be considered within the context of environmental responses and consequences. The questions of causal process and controlling variables (soil character, vegetation cover, gradient, precipitation and moisture characteristics) are extensively covered in the existing literature and are not at issue here. What are fundamentally different are the consequences. In developing regions, soil erosion and the associated environmental consequences directly affect the livelihood of the inhabitants, frequently resulting in starvation and even death of local inhabitants. By contrast, the worst consequences in highly developed nations are slight increases in national unemployment figures (Rich, 1994) and reductions in export revenues (Sidle *et al.*, 1985). Such consideration is of the utmost importance for Africa, where it has been estimated that more than 60% (and in some regions up to 80%) of the population are still directly or indirectly reliant upon adequate agricultural yields. It is evident from the work of Garland (1990) and Beckedahl (1998) that this dichotomy is not only relevant when comparing nations, but also exists among regions within African countries themselves.

It is important that the socio-economic implications of soil erosion be viewed within the context of the discourse presented by Scoones *et al.* (1996), who acknowledge that soil erosion in Africa constitutes a major problem, but warn against what has been labelled a 'crisis narrative' (Roe, 1995) – such rhetoric is akin to a Malthusian-type apocalypse precipitated by environmental collapse consequent upon severe degradation and used by some developmentalists to claim stewardship over land and resources they do not own. Such considerations led to the argument that, while erosion is undoubtedly a problem undermining agricultural production in parts of Africa (see Fig. 6.1), there is also a need to recognize the existence of other important production constraints (Stocking, 1995, 1996; Scoones *et al.*, 1996). This line of enquiry moves the debate away from generalizations and towards a scientific analysis of the nature and causality of soil erosion in Africa, as well as its implications and consequences. Such argument should also include due consideration of indigenous soil conservation techniques; a recognition which is driven from the realization that, in order to attain soil conservation, the techniques must gain acceptance by the people directly affected (see Tiffen *et al.*, 1994; Sikana and Mwambazi, 1996).

Fig. 6.1. 'Badland' topography developed on degraded agricultural land in the Tsolo district of the Eastern Cape Province, South Africa.

The Biophysical Context of Soil Erosion in Africa

Separating short-term, human-induced erosion from longer-term natural erosion processes is as important to soil conservation as issues related to affected communities (e.g., Blaikie, 1985; Wells and Andriamihaja, 1993; Christiansson and Kikula, 1996; Beckedahl, 1998). Africa, being situated approximately symmetrically with respect to latitude, is extensively influenced by a heat exchange driven global atmospheric circulation and can thus play a crucial role in revealing possible causal relationships between climate change, soil erosion and socio-economic consequences (Tyson, 1986; Meadows, 1988; Goudie, 1996; Anhuf *et al.*, 1999). Of possible equal importance is the synergistic role played by tectonic activity and the associated ruggedness of parts of the continent (e.g., Rapp *et al.*, 1972; Cooper, 1990).

The problem of contemporary erosion rates that exceed estimated rates of soil formation has already been alluded to. As pertinent is the observation by Beckedahl *et al.* (1988) in South Africa, Shakesby and Whitlow (1991) in Zimbabwe, and Bryan (1994) in Kenya that the rates of natural (or geologically normal) erosion in Africa may, at least in places, exceed the rate of soil formation by a factor of more than two. It is well established that erosion severity is affected by antecedent moisture conditions and vegetation cover, and is further influenced by both the magnitude and frequency of precipitation events as well as by the erodibility of the soil. It is here that another major difference exists between erosion in Africa and elsewhere.

The landscapes of Africa have been largely unaffected by continental glaciation since the breakup of the Gondwana Pangea some 180 million years ago (Tankard *et al.,* 1982). In contrast to most soils of Europe and North America that have developed since the last Ice Age, many soils of Africa have been developing since at least the Miocene (Dingle *et al.,* 1983) and are hence generally well-leached, skeletal soils that tend to be erodible. Much work on the question of soil erosion in Africa has, for a variety of reasons, relied on extensive use of the USLE (i.e., Wischmeier and Smith, 1978), not withstanding the cautionary note of Bergsma and Valenzuela (1981). It has been argued that African soils may be broadly assigned to one of six erodibility classes as discussed by Nill *et al.* (1996). Another empirical model that has to some extent been favoured in southern Africa is SLEMSA (Soil Loss Estimator for Southern Africa) (Schulze, 1979; Elwell, 1984). The extensive use of these empirical models can be attributed to the fact that they provide a 'first approximation' of the problem and are cost effective even if not highly accurate. A similar argument can probably be invoked to explain the reluctance to recalculate soil loss based on the more physically-based erosion models such as RUSLE (Revised Universal Soil Loss Equation; e.g., Renard *et al.*, 1997) and WEPP (Watershed Erosion Prediction Project; e.g., Laflen *et al.,* 1991). Many of the parameters used in the empirical models are, in the African context, estimates and hence no tangible improvement in the predictive accuracy of these more sophisticated models is likely without extensive further research.

The other landscape parameter that needs to be considered in Africa is the 'slope factor' – i.e., the rugged nature of the terrain coupled with the length of slope. Although the slope length can be artificially modified by conservation measures, it is considerably more costly to address the question of gradient (although this has been done successfully on a local scale in many African countries). Slope plays an important role through its effect on natural erosion, however, many African countries (e.g., Lesotho) are reliant upon steep land agriculture. Associated high soil losses are then further exacerbated by the land use practices linked to population pressure (Bojo, 1991). Certain agroforestry practices such as small terraces, grass strips, and hedgerows offer protection against soil and nutrient losses from cultivated hillslopes (Roose and Ndayizigiye, 1997; Tamubula and Sinden, 2000).

Climate and related vegetation cover are the other 'natural' parameters affecting soil loss. Those regions of Africa experiencing seasonal or episodic rainfall (most of Africa outside of the tropics) are prone to large-scale variations in antecedent soil moisture and are partly susceptible to dessication and trampling with the associated risk of the destruction of soil structure. The marked seasonality of the precipitation in many of these areas results in a period of dormancy or even die-back in the vegetation. Often the first rains of the season have high intensities, thus making the soil prone to severe erosion (Fig. 6.2) (Stocking, 1981).

Fig. 6.2. Severe erosion in the Kutsolo area, Maclear District, Eastern Cape Province, South Africa.

Soil Loss in the Context of African Environmental Change

It is important to review the effect of projected climate-change scenarios on soil erosion potential in Africa. At present, the most reliable predictions appear to be derived from analyses of paleoenvironmental conditions (e.g., Tyson, 1986; Deacon and Lancaster, 1988; Littmann, 1988; Meadows, 1988; Anhuf *et al.*, 1999). Goudie (1996), building upon the research of Griffiths (1987), noted that it is possible to broadly divide Africa into eight major climatic regions on the basis of precipitation and temperature. Each of these zones or regions is relatively distinctive in terms of climatic variability and rainfall reliability. Analysis of Holocene and more recent climatic fluctuations reveals that climate anomalies of a magnitude required to produce past environmental conditions are not uncommon today, and that 'the long-term changes of environment during the past millennium appear to represent primarily a more persistent occurrence of conditions commonly characterizing briefer periods of the twentieth century' (Nicholson, 1996; p. 84).

The present favoured scenario derived from Global Climate Models (GCMs) appears to support Nicholson's (1996) observation, suggesting little change in total precipitation, but an increase in larger magnitude events. Such a scenario implies that both rainfall erosivity and dessication (with the associated risk of trampling) will increase, hence increasing soil loss (Table 6.1). However, if global warming occurs at predicted rates, anthropogenic changes will likely be induced as well, further worsening the situation. Land use practices will then need to change to accommodate the decreased water availability. Such responses are likely to place greater pressures on the land, enhancing degradation and gullying (Fig. 6.3). This degradation could stabilize relatively rapidly if conservation farming and water harvesting techniques are adopted.

Fig. 6.3. Severe gullying in the Ncise area near Umtata, Eastern Cape Province, South Africa, resulting from a combination of environmental and anthropogenic factors.

Table 6.1. The probable erosional consequences of global warming for Africa and potential remedial measures which can be taken to minimize these effects (broken lines indicate related processes).

Erosion process	Projected change due to global warming effects	Potential remedial measures
Rain-splash erosion	increase due to higher rainfall energy	increase mulch and groundcover
Wind erosion	increase due to increase in dessication and decrease of soil moisture due to more high-magnitude low-frequency events	increase windbreaks and mulch to reduce wind energy and increase soil moisture
Airborne dust and dust deposition	increase resulting from increased wind erosion and the effects of trampling	increase trapping efficiency and lower wind energy by increasing windbreaks
Sheet erosion	increase due to increased availablity of material as a result of trampling and more free water associated with high-intensity rain	increase matted vegetation and mulch; avoid overgrazing to reduce risk of trampling
Down-slope / off-site siltation	increase resulting from increased sediment load	increase surface roughness and mulch to promote infiltration and reduce overland flow speed
Rills and gullies	increase due to projected increase in free surface water resulting from high-intensity precipitation	increase groundcover; judicious ploughing / conservation tillage to maintain soil structure and root cohesion of soil: increase in surface roughness e.g. berms and rainwater harvesting
Siltation of rivers and reservoirs	probable increase due to projected significant increase in sediment transport across the land and ultimately into the river systems	meticulous attention to conservation practices in all facets of land husbandry; establish and conserve wetlands within river systems to act as natural sediment sinks
Subsurface erosion / soil piping and land degradation	likely to increase in degraded soils due to increased dessication cracking promoting selective infiltration into the soil profile	increase mulching and groundcover, including conservation tillage to limit dessication cracking; rainwater harvesting

Further complicating factors in many African countries are the socio-politically induced land-use changes associated with land reforms that are at present concentrated largely in southern Africa. Such reforms will increase the rural population, increase subsistence agriculture and potentially decrease the commercial agricultural sector throughout much of Africa. An indirect consequence is a potential net decrease in absolute poverty for individual countries yet creating an agricultural sector which is less well equipped to cope, both materially and educationally, with the causes and consequences of soil erosion. Although for many Africans the ideals of land reform initiatives are viewed as a return to the land from which they moved under duress, the reality is that most people affected are not experienced small-landholding agriculturalists. The risk thus exists that the reforms designed to rectify past injustices, while simultaneously alleviating some of the side effects of unemployment, may well place significant areas of marginal land at risk. Whether or not the African land reform initiatives will result in actually increasing soil erosion rates depends on available resources and on the effectiveness of information dissemination to the affected communities.

Further uncertainty arises for the arid and semiarid areas already operating under drought stress. Here it is the question of extent that will determine the land use potential and the resultant soil loss. The unfortunate reality for many subsistence farmers is that they have little choice but to cultivate their fields to the best of their ability given the limited resources at their disposal, and to accept high soil losses. The International Centre for Research in Agroforestry (ICRAF) located in Nairobi, Kenya, has become a focal point for promoting agroforestry systems designed to improve the nutritional, economic, and social well being of people of developing countries as well as to achieve more sustainable land use. Recent studies of treecrop interactions in such semiarid systems have yielded promising results in terms of soil erosion protection and nutrient and water conservation (e.g., Ong *et al.*, 2000); however, much additional work is needed to develop agroforestry systems that are both appropriate and easy to implement in Africa.

In considering the potential consequences of global warming on soil erosion, it is of critical importance not to oversimplify the complexity of the land-soil-climate system by focusing too closely upon direct dependencies and ignoring the potential for buffer effects caused by feedback mechanisms within the system. Scharpenseel and Pfeiffer (1998) suggest that such effects may impede the projected effects of the Gulf Stream-related temperature rise. Their work does, however, also show that the ENSO (El Ninõ Southern Oscillation) climatic extremes contribute to the catastrophic destruction of soil - a situation akin to what is potentially anticipated under the global warming scenario for southern Africa. The upper few centimetres of soil are believed to be most affected by climatic conditions but are also the most vulnerable to tillage and land-use practice (Kimble *et al.*, 1998).

The Effect of Population on Soil Loss in Africa

Table 6.2. Conceptual representation of the potential interdependence between demographic and socio-economic parameters and soil erosion in Africa.

Scenario	Consequences for population	Agricultural implications	Likely implications for soil erosion rates
High population growth rates	Increase in both rural and urban population	Increased pressure on land for production in both the commercial and subsistence agricultural sectors	Increase
Land reform / redistribution	Increase in rural population	Increased subsistence and smallholder agriculture	Increase
High mortality due to AIDS pandemic (low population growth rates)	Decrease in population	Less *per capita* pressure on the land	Decrease
	Decrease productivity of population due to poor health; decrease in available labour	Decrease in the potential for conservation farming	Increase
Changing global economy due to climatic change	Increase unemployment due to adverse climate	Decrease in total resources available for conservation farmer, especially at the subsistence and smallholder levels	Increase
	Decreased unemployment and increased yields due to a more favourable climate	Increase in total resources available for conservation practices	Decrease

As discussed earlier, erosion and the potentially diminished agricultural yield are only part of the greater social, political and economic issues in Africa (Blaikie, 1985; Scoones *et al.,* 1996). As population and especially rural population increases in Africa, the pressure on the land will increase with the associated increased likelihood for erosion. The uncertainty, however, revolves around the nature and extent of the environmental changes, coupled with land-use change (e.g., Simmons and Mannion, 1995; Christiansson and Kikula, 1996) as well as the population dynamics – the degree of urbanization and the effect of the AIDS pandemic (Table 6.2). Increased afforestation is one of the primary land-use changes that would promote soil conservation and resource sustainability in rural Africa (e.g., König, 1992; Gubbels, 1994). Additionally, advanced and applicable agroforestry practices offer both socio-economic and biophysical benefits related to resource sustainability and soil conservation, however, little information exists on the impacts of such dispersed systems at larger-scales (Steiner, 1988; Fischer and Vasseur, 2000).

Conclusions

Soil erosion rates in Africa are highly variable, from as low as 1.5 t ha^{-1} yr^{-1} to an excess of 250 t ha^{-1} yr^{-1}. It is unlikely that this large variance will change in the near future as a result of climate change. What will change is the rate of soil loss for specific regions and districts. Erosion rates in Africa in the future are likely to reflect demographic factors even more strongly than before. The nature and extent of changes in erosion rates will reflect the awareness of people living in these areas to the inherent risks and whether this impacted population will have the human and material resources to implement effective soil conservation measures.

Thus the role of humans in environmental degradation appears to be two-fold: (i) their role as principal factors or contributors of global change; and (ii) land misuse caused by humans. Together these factors may contribute to a progressive downward spiral of environmental degradation.

References

Anhuf, D., Frankenberg, P. and Lauer, W. (1999) Die Postglaziale Warmphase vor 8000 Jahren – Eine Vegetationsrekonstruktion für Afrika. *Geographische Rundschau* 51, 454-461.

Beckedahl, H.R., Bowyer-Bower, T.A.S., Dardis, G.F. and Hanvey, P.M. (1988) Geomorphic effects of soil erosion. In: Moon, B.P. and Dardis, G.F. (eds), *Geomorphology of Southern Africa*. Southern Books, Johannesburg, South Africa, pp. 249-276.

Beckedahl, H.R. (1998) Subsurface Soil Erosion Phenomena in South Africa. *Petermanns Geographische Mitteilungen Monograph Series* 290, Justus Perthes Verlag, Gotha.

Bergsma, E. and Valenzuela, C.R. (1981) Drop testing aggregate stability of some soils near Mérida, Spain. *Earth Surface Processes and Landforms* 6, 309-318.

Blaikie, P. (1985) *The Political Economy of Soil Erosion in Developing Countries.* Longman, London.

Bojo, J. (1991) *The Economics of Land Degradation: Theory and Application in Lesotho.* Stockholm Environment Institute, Stockholm.

Bork, H.R. (1988) Bodenerosion und Umwelt. *Landschaftsgenese und Landschaftsökologie.* Heft 13, Technische Universität Braunschweig, Germany.

Boucher, K. and Weaver, A.B. (1991) Sediment yield in South Africa - a preliminary geographical analysis. *GeoJournal* 23, 7-17.

Brown, L.R., Durning, A., Flavin, C., French, M., Jacobson, J., Lowe, M., Postel, S., Renner, M., Starke, L. and Young, J. (1990) *State of the World.* Worldwatch Institute and W.W. Norton, New York.

Bryan, R.B. (ed.) (1994) Soil Erosion, Land Degradation and Social Transition - Geoecological analysis of a semi-arid tropical region, Kenya. *Catena Supplementary Series* 27, Catena Verlag, Cremlingen-Destedt, Germany.

Christiansson, C. and Kikula, I.S. (eds) (1996) Changing Environments: Research on Man-Land Interrelations in Semi-arid Tanzania. *Regional Soil Conservation Unit Report 13.* SIDA, Nairobi.

Cooper, M.R. (1990) Tectonic cycles in southern Africa. *Earth Science Reviews* 28, 321-364.

Deacon, J. and Lancaster, N. (1988) *Late Quaternary Palaeoenvironments of Southern Africa.* Oxford University Press, Oxford, UK.

Dingle, R.V., Siesser, W.G. and Newton, R.A. (1983) *Mesozoic and Tertiary Geology of Southern Africa.* Balkema, Rotterdam.

Elwell, H.A. (1984) Soil loss estimation: a modelling technique, In: Hadley, R.F. and Walling, D.E. (eds), *Erosion and Sediment Yield - Some Methods of Measurement and Modelling.* University Press, Cambridge, UK, pp. 15-30.

FAO (1983) Keeping the land alive - its causes and cures. *Soils Bulletin* 50. FAO, United Nations, Rome, Italy.

Fischer, A. and Vasseur, L. (2000) The crisis in shifting cultivation practices and the promise of agroforestry: a review of the Panamanian experience. *Biodiversity and Conservation* 9, 739-756.

Garland, G. (1990) The soil spoilers. In: Preston-Whyte, R.A. (ed.), *Rotating the Cube: Environmental Strategies for the 1990's.* University of Natal, Durban.

Goudie, A.S. (1996) Climate: Past and present. In: Adams, W.M., Goudie, A.S. and Orme, A.R. (eds), *The Physical Geography of Africa.* Oxford University Press, Oxford, pp. 34-59.

Griffiths, J.F. (1987) Climate of Africa. In: Oliver, J.E. and Fairbridge, R.W. (eds), *The Encyclopedia of Climatology.* Harper, New York, pp. 572-589.

Gubbels, P. (1994) Populist pipe-dream or practical paradigm? Farmer driven research and the project agroforestier. In: Scoones, I. and Thompson, J. (eds), *Beyond Farmer First: Rural People's Knowledge, Agricultural Research and Extension Practice.* IT Publications, London, pp. 238-243.

Kimble, J.M., Lal, R. and Grossman, R.B. (1998) Alteration of soil properties caused by climate change, In: Blume, H.P., Eger, H., Fleischhauer, E., Hebel, A., Reij, C. and Steiner, K.G. (eds), Towards sustainable land use. *Advances in GeoEcology* 31, 175-184.

König, D. (1992) The potential of agroforestry methods for erosion control in Rwanda. *Soil Technology* 5, 167-176.

Laflen, J.M., Elliot, W.J., Simanton, R., Holzhey, S. and Kohl, K.D. (1991) WEPP soil erodibility experiments for rangeland and cropland soils. *Journal of Soil and Water Conservation* 46, 39-44.
Littmann, T. (1988) *Jungquartäre Ökosystemveränderungen und Klimaschwankungen in den Trockengebieten Amerikas und Afrikas.* Ferdinand Schöningh, Paderborn, Germany.
Meadows, M.E. (1988) Landforms and Quaternary climatic change. In: Moon, B.P. and Dardis, G.F. (eds), *Geomorphology of Southern Africa.* Southern Books, Johannesburg, South Africa, pp. 296-316.
Middleton, N. and Thomas, D. (1997) *World Atlas of Desertification.* 2nd edition. UNDP, Arnold Publ., London.
Millington, A.C. (1992) Soil Erosion and Conservation. In: Mannion, A.M. and Boulby, S.R. (eds), *Environmental Issues in the 1990's.* John Wiley and Sons, Chichester, pp. 227-244.
Nicholson, S.E. (1996) Environmental change within the Historical Period. In: Adams, W.M., Goudie, A.S. and Orme, A.R. (eds), *The Physical Geography of Africa.* Oxford University Press, Oxford, pp. 60-87.
Nill, D., Schwertmann, U., Sabel-Koschella, U., Bernhard, M. and Breuer, J. (1996) *Soil Erosion by Water in Africa - Principles, Prediction and Protection.* GTZ Volume 257, GTZ, Eschborn, Germany.
Ong, C.K., Black, C.R., Wallace, J.S., Khan, A.A.H., Lott, J.E., Jackson, N.A., Howard, S.B. and Smith, D.M. (2000) Productivity, microclimate and water use in *Grevillea robusta*-based agroforestry systems on hillslopes in semi-arid Kenya. *Agriculture Ecosystems and Environment* 80, 121-141.
Pimentel, D., Harvey, C., Resosudarmo, P., Sinclair, K., Kurz, D., McNair, M., Crist, S., Shpritz, L., Fitton, L., Saffouri, R. and Blair, R. (1995) Environmental and economic costs of soil erosion and conservation benefits. *Science* 267, 1117-1123.
Rapp, A., Murray-Rust, D.H., Christiansson, C. and Berry, L. (1972) Soil erosion and sedimentation in four catchments near Dodoma, Tanzania. *Geografiska Annaler* 74A, 255-318.
Renard, K.G., Foster, G.R., Weesies, G.A., McCool, D.K. and Yoder, D.C. (1997) Predicting Soil Erosion by water: A Guide to Conservation Planning with the Revised Universal Soil Loss Equation (RUSLE). *USDA Handbook 703*, USDA-ARS, Tuscon, Arizona.
Rich, B. (1994) *Mortgaging the Earth.* Earthscan Publications, London.
Roe, E. (1995) Except Africa: Postscript to a special section in development narratives. *World Development* 19, 287-300.
Roose, E. and Ndayizigiye, F. (1997) Agroforestry, water and soil fertility management to fight erosion in tropical mountains of Rwanda. *Soil Technology* 11, 109-119.
Scharpenseel, H.-W. and Pfeiffer, E.M. (1998) Impacts of possible climate change upon soils; some regional consequences. In: Blume, H.-P., Eger, H., Fleischhauer, E., Hebel, A., Reij, C. and Steiner, K.G. (eds), Towards sustainable land use. *Advances in GeoEcology* 31, 193-208.
Schulze, R.E. (1979) Soil loss in the key area of the Drakensberg: A regional application of the Soil Loss Estimation Model for Southern Africa (SLEMSA). *Agricultural Engineering in South Africa* 232, 22-33.
Scoones, I., Reij, C. and Toulmin, C. (1996) Sustaining the soil: Indigenous soil and water conservation in Africa. In: Reij, C., Scoones, I. and Toulmin, C. (eds), *Sustaining the Soil - Indigenous Soil and Water Conservation in Africa.* Earthscan Publications, London, pp. 1-27.

Shakesby, R.A. and Whitlow, R. (1991) Perspectives on prehistoric and recent gullying in central Zimbabwe. *GeoJournal* 23, 49-58.

Sidle, R.C., Pearce, A.J. and O'Loughlin, C.L. (1985) *Hillslope Stability and Land Use.* American Geophysical Union, Water Resources Monograph No. 11, Washington, DC.

Sikana, P.M. and Mwambazi, T.N. (1996) Environmental change and livelihood responses: shifting agricultural practices in the lakes depression zone of Northern Zambia. In: Reij, C., Scoones, I. and Toulmin, C. (eds), *Sustaining the Soil - Indigenous Soil and Water Conservation in Africa.* Earthscan Publications, London, pp. 107-116.

Simmons, I.G. and Mannion, A.M. (eds) (1995) The Changing Nature of the People - Environment Relationship: Evidence from a Variety of Archives. In: *Proceedings of the International Geographical Union Commission on Historical Monitoring of Environmental Changes Meeting, August 18–21 1994, Charles University, Praha/Prague, Czech Republic.*

Steiner, F. (1988) Agroforestry's coming of age. *Journal of Soil and Water Conservation* 43, 157-158.

Stocking, M.A. (1981) Conservation strategies for less developed countries. In: Morgan, R.P.C. (ed.), *Soil Conservation: Problems and Prospects.* John Wiley and Sons, Chichester, UK, pp. 377-383.

Stocking, M.A. (1984) Rates of erosion and sediment yield in the African environment. In: Walling, D.E., Foster, S.S.D. and Wurzel, P. (eds), *Challenges in African Hydrology and Water Resources, IAHS Publication* 144, 285-294.

Stocking, M.A. (1995) Soil erosion in developing countries: Where geomorphology fears to tread. *Catena* 25, 253-267.

Stocking, M.A. (1996) Soil erosion. In: Adams, W.M., Goudie, A.S. and Orme, A.R. (eds), *The Physical Geography of Africa.* Oxford University Press, Oxford, UK, pp. 326-341.

Tamubula, I. and Sinden, J.A. (2000) Sustainability and economic efficiency of agroforestry systems in Embu District, Kenya: an application of environmental modelling. *Environmental Modelling and Software* 15, 13-21.

Tankard, A.J., Jackson, M.P.A., Eriksson, K.A., Hobday, D.K., Hunter, D.R. and Minter, W.E.L. (1982) *Crustal Evolution of Southern Africa: 38 Billion years of Earth History.* Springer Verlag, New York.

Tiffen, M., Mortimore, M. and Gichuki, F. (1994) *More People, Less Erosion: Environmental Recovery in Kenya.* John Wiley, Chichester, UK.

Tyson, P.D. (1986) *Climatic Change and Variability in Southern Africa.* Oxford University Press, Cape Town, South Africa.

Wells, N.A. and Andriamihaja, B. (1993) The initiation and growth of gullies in Madagascar: are humans to blame? *Geomorphology* 8, 1-46.

Wischmeier, W.H. and Smith, D.D. (1978) Predicting rainfall erosion losses. *USDA Agricultural Research Services Handbook 537.*

Chapter 7

The Effects of Environmental Changes on Weathering, Gravitational Rock Deformation, and Landslides

Masahiro Chigira
Disaster Prevention Research Institute, Kyoto University, Japan

The effects of climatic change on the processes of weathering, slow and long-term gravitational deformation of rocks, and rapid landslides are discussed. Weathering processes, which result in landslide materials, create weathering profiles during variable time periods ranging from tens of years to geologic time. Additionally, weathering determines the location of potential slip surfaces or landslide behaviour. However, the characteristics of weathering profiles are poorly understood. Some types of rocks are weathered rapidly and form weathering profiles that favour landsliding; slopes composed of those rocks are affected by environmental change through weathering. Unwelded ignimbrite is one of the most vulnerable rocks with recurrence intervals of landsliding less than a few hundred years. Other vulnerable rocks include marine mudstones and sandstones. Weathered granite is also vulnerable to weathering because it is reweathered when it is exposed by erosion. Reweathering of granite results in an abrupt weathering front that forms the slip surface during heavy rains. Gravitational deformation of a rock mass (mass rock creep) proceeds gradually but continuously, sometimes concentrating at a specific depth but other times spreading over various depths beneath a slope. The distributed deformation creates many fractures, eventually changing the rock mass to rock debris; the debris is supported by plant roots but may fail if the support is lost after logging, wildfire, or environmental changes, or with further deformation. Changes in groundwater conditions affect the stability of rock slopes. Groundwater that drains from a rock mass under a temperate climate would be locked in if the temperature drops and the resultant freezing of the rock surface creates a membrane surrounding the rock mass. Such a membrane effect could generate a rockfall on a steep rock cliff because of the increase in bulk weight.

Introduction

The opposing forces of uplift and denudation dynamically balance mountain slopes. Denudation, which occurs through the integrated processes of weathering, mass movement, and surface erosion, is affected by environmental changes, particularly climatic change. Climatic change affects the timing and rates of rock weathering, groundwater levels, and vegetation by altering precipitation, evapotranspiration, and temperature magnitudes and regimes. These changes result in the destabilization or stabilization of slopes as discussed by the TESLEC project (temporal stability and activity of landslides in Europe with respect to climatic change) within the framework of the Environmental Programme of the Commission of the European Communities (Schrott and Pasuto, 1999). This chapter summarizes the effects of climatic change on the processes of weathering, slow and long-term gravitational deformation of rocks, and rapid landslides.

Weathering proceeds from the earth surface and thickens weathering profiles depending on rock type and climatic conditions. Along with groundwater conditions, weathering profiles often dictate the spatial characteristics of shallow landslides because their structure typically determines the location of slip surfaces and the probability of slope failure (Chigira, 2001; Chigira and Inokuchi, in press). The development of weathering profiles decreases slope stability and eventually leads to landslides. Weathering is accelerated by increases in rainfall, which also decreases slope stability by increasing the pore pressure of groundwater. A change in rainfall intensity over a long period thus affects the weathering conditions of rocks and also lowers the triggering threshold for landslide initiation as will be described later.

Aside from the surface weathering process, the gravitational deformation of rocks proceeds slowly but continuously and provides the basic preconditions for rapid and catastrophic landslides, the occurrence of which are also sensitive to environmental changes. Rock masses are gradually deformed within rock slopes; when subjected to a gravitationally unstable state for a long time, they form various types of deformed structures and debris (Chigira, 1992). The resultant loosened surface materials or debris are easily affected by environmental changes. On the other hand, a rock mass with a zone of deformation at great depth is not influenced by environmental change directly, but would slide due to an ascending groundwater level or undercutting by erosion.

Development of Weathering Zones

Previous studies on weathering in relation to landslides

Weathering has been studied in various fields of science, such as geology, engineering geology, geochemistry, mineralogy, geomorphology, and soil science, but the relationship between weathering and the occurrence of landslides has not been well understood as yet because of inadequate knowledge of weathering profiles. A weathering profile generally consists of solum (topsoil, O, A, and B horizons), saprolite, and partially weathered rocks (Selby, 1993), of which only topsoil has been studied well by soil scientists. However, landslides often involve materials below topsoil, and the formative processes of these materials have not been sufficiently clarified. Geologists and engineering geologists have investigated weathering profiles related to engineering-geological projects (Moye, 1955; Ruxton and Berry, 1957; Dearman, 1974), but recent publications in scientific journals are few. On the other hand, in the 1970s geochemists initiated studies on chemical weathering with respect to water-rock interactions and made significant advances in the studies of reaction kinetics, mineral-surface chemistry and chemical budgets (Gardner, 1980; Fritz, 1988; Hyman *et al.*, 1998; White and Brantley, 1995). However, this research focused on the reactions between minerals and water rather than on reactions within a weathering profile. Focus on the latter is essential for characterizing the occurrence of weathering profiles and understanding their formative mechanisms (White *et al.*, 1998). Clay mineralogists have studied the relationship between mineralogy and weathering for many years. Geomorphologists have also investigated weathering processes (Yatsu, 1988; Matsukura, 1989; Selby, 1993), but their major interests relate to the earth surface rather than to subsurface weathering profiles.

Weathering profiles in forested areas are formed mainly by chemical and biological weathering processes, while some types of physical processes, particularly the sequence of drying and wetting, may influence the mechanical properties of shallow parts of weathering profiles. Other physical processes, such as stress release and the interaction of freezing and thawing, may not be as important in forested areas compared to agricultural or developed landscapes. Sheeting may occur to a depth of 100 m, but does not form weathering profiles by itself (Folk and Patton, 1982; Holzhausen, 1989), whereas micro-sheeting joints may form weathering profiles (see later discussion; Chigira, 2001).

The weathering rate, which is important for pre-conditioning landslide materials, depends upon rock types, structures, hydraulic conditions, and climate. The weathering rate thus is very sensitive to environmental change. Common hard rocks are resistant to weathering on a time scale of tens to hundreds of years, as evidenced by the endurance of many historic stone monuments. Therefore, slopes developed from fresh resistant rocks do not become unstable as the result of weathering in short periods. However, other types of rocks, such as weakly consolidated sedimentary rocks and soft pyroclastics, are easily weathered (either chemically or physically) and are therefore prone to landslides.

Chemical weathering accelerates with increases in both rainfall and consequent infiltration into groundwater because reactive water is continuously supplied through the vadose zone and saturated substrate. The chemical weathering rate is also enhanced by increases in temperature, as is generally well known by the Arrhenius equation for chemical reactions. By studying the weathering of gravels from precisely dated terrace deposits, Nishiyama *et al.* (1999) concluded that chemical weathering was more rapid during the last interglacial age compared with more recent historical records. This rapid weathering may be due to higher temperatures or a more humid climate.

In addition to rainfall intensity, the pH of water also influences chemical weathering. Since the industrial revolution, acidic water from acid rain has greatly deteriorated stone structures in Europe (Winkler, 1994). Acidic water can also be generated underground by chemical reactions, such as through the oxidation of sulphide minerals. In addition, chemical dissolution rates of silicate minerals depend on the pH of water and increase dramatically when pH falls below 4 or 5 (Drever, 1997), which are levels that acid rain can reach.

Chemical weathering, particularly the interaction between carbonic acid and minerals, has been debated as a possible factor related to climatic change. For example, the Himalayan uplift has been implicated as a major factor in the global climatic cooling of the past 40 million years because of increased silicate weathering and consequent increases in atmospheric CO_2 consumption (Raymo and Ruddiman, 1992). These authors note that the uplift generated large amounts of debris of silicate minerals that are comminuted by physical processes to yield large specific surfaces, resulting in fast chemical reactions. This idea was supported by dramatic increases in marine $^{87}Sr/^{86}Sr$ ratios at 40 million years ago, just after the commencement of the Asia-India continental collision. However, Blum *et al.* (1998) insisted that the increase in isotope ratio is due to the dissolution of trace amounts of calcite within the dominantly silicate High Himalayan Crystalline rocks, and that Raymo and Ruddiman (1992) overestimated the amount of CO_2 consumption due to silicate weathering.

Characteristic weathering profiles and landslides

Ignimbrite

Pyroclastics, such as ash and tuff, are easily weathered by hydration, ion exchange, dissolution, and clay mineralization (Petit *et al.*, 1990). Pyroclastics are widely distributed in regions with volcanism, such as Japan (Cas and Wright, 1996). Pyroclastic flow deposits (i.e., ignimbrite), one of the most common pyroclastics, may be subject to landslides during heavy rains.

Shirasu, a typical unwelded ignimbrite (age 25.000 years before present), has been subjected to shallow landslides on many occasions resulting in numerous casualties in Kagoshima Prefecture, southern Japan (Fig. 7.1) (Yokota, 1999). Shirasu is weathered so rapidly that, after a landslide strips away weathered material, weathering commences again and quickly provides material for the next landslide (Fig. 7.2). The weathering intensity of Shirasu is strongest at the ground surface and decreases with depth. The strength of Shirasu, as measured by indices obtained from portable cone penetration tests or cone penetrometers, decreases linearly from the ground surface to the depth of the weathering front, thereafter being constant (Shimokawa *et al.*, 1989; Yokota, 1999). The migration of a weathering front and the deterioration within the weathering zone proceed on the order of years, and the recurrence interval of landslides is on the order of tens of years to a few hundred years (Shimokawa *et al.*, 1989). The weathering mechanism has not been sufficiently elucidated but may be dominated by hydrogeochemical interactions between the Shirasu deposits and groundwater as indicated by the increase in water content toward the ground surface. Therefore, changes in precipitation, particularly from dry to humid conditions, would accelerate weathering and consequently increase the occurrence of landslides.

Fig. 7.1. Landslides of unwelded pyroclastic flow, Shirasu, Kyushu, Japan. (Photo by Kokusai Koku Photo.)

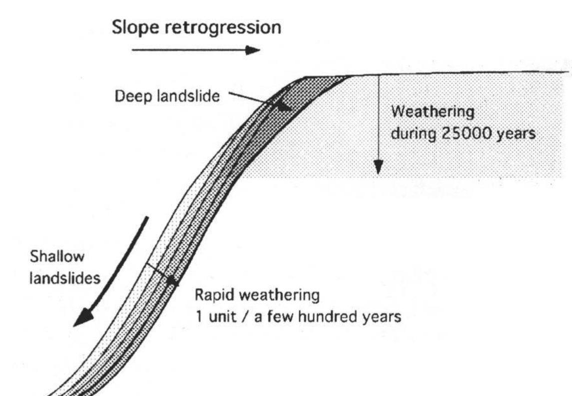


Fig. 7.2. Schematic sketch showing the iteration of landslides and the weathering of an unwelded pyroclastic flow. (Modified from Shimokawa *et al.*, 1989.)

Unwelded ignimbrite or pumiceous pyroclastic material underlies Kozu Island, south of Tokyo. An earthquake of magnitude 6.4 occurred in this area on 1 July 2000. A rainstorm with a daily precipitation of 293 mm followed, resulting in many shallow landslides (Fig. 7.3). Preliminary investigations indicated that this pyroclastic deposit has a weathering profile similar to that of Shirasu and the weathered zone's surface layer, which was less than 1 m thick, slid in many locations (Chigira, unpublished data). Repeated sequences of weathering and shallow landsliding also appear to have occurred in this area.

Fig. 7.3. Landslides of acidic pyroclastics generated by earthquakes and rainfall in Kozu Island, south of Tokyo, Japan

Weakly welded ignimbrite or vapour-phase crystallized ignimbrite is not as easily weathered as unwelded ignimbrite, but it commonly has a very distinct weathering profile. Such profiles were recognized as the basic causes of landslides during the heavy rains of 1999 in Fukushima Prefecture, in northeastern Japan (Chigira and Inokuchi, in press). The ignimbrite, Quaternary Shirakawa Pyroclastic Flow, is generally separated into thin plates parallel to the ground surface at the base of the heavily weathered portion of the deposit beneath the topsoil (Fig. 7.4A). This zone, called a foliated zone, is 30 cm to about 1 m thick, each plate ranging from a few to 5 cm in thickness. The tuff in the foliated zone and the overlying soil are much softer than the underlying tuff, although few clay minerals have been detected in the rocks from this zone and the overlying soil. According to the structure of the weathering zone, rainwater infiltrates into the soil and the foliated zone along fractures, however the water cannot penetrate the underlying weakly weathered and unweathered rocks because of the scarcity of cracks. Consequently, the soil and the rock of the foliated zone rapidly become saturated, leading to sliding

(Fig. 7.4B). Because of the clearly defined weathering front and the absence of cracks beneath the foliated zone, plant roots do not penetrate downward from the foliated zone to support the surface materials of weathered tuff and colluvium. The exfoliation of the ignimbrite was generally observed beneath the slopes that cut the top of the Shirakawa pyroclastic flow deposits and also cut the overlying tephra for 20,000 to 30,000 years (Suzuki, 1992). The mechanism of exfoliation has not yet been clarified.

Fig. 7.4A. Foliated zone in the weathering profile of vapour-phase crystallized ignimbrite.

Fig. 7.4B. Landslide of ignimbrite of the Shirakawa Pyroclastic Flow.

Non-welded or weakly welded ignimbrite tends to be prone to landslides as described previously, but strongly welded ignimbrite generally is not susceptible to landslides, except for the occasional toppling of rock columns divided by columnar joints. Therefore, landslide hazard assessment must consider the degree of welding in such deposits.

Marine sedimentary rocks

Marine sedimentary rocks, particularly soft sandstone and mudstone, are vulnerable to rapid weathering, as is indicated by Chigira (1990), Pye and Miller (1990), Chigira and Sone (1991), Chigira and Oyama (1999), and Oyama and Chigi- ra (1999). The most significant weathering of soft marine

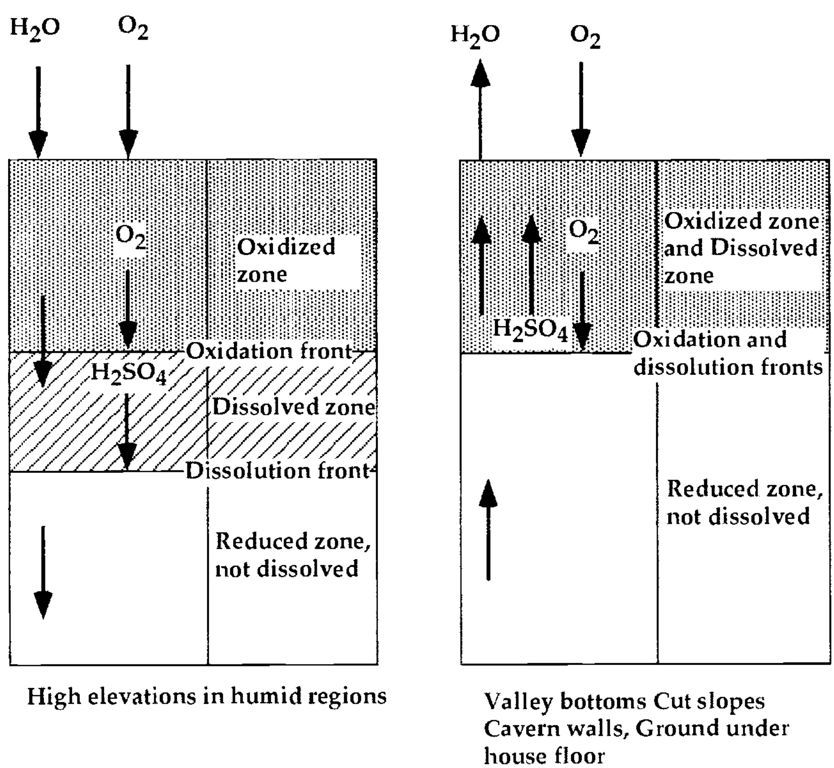


Fig. 7.5. Characteristic weathering profile of marine sedimentary rocks (Chigira and Oyama, 1999). When the fluxes of oxygen and sulphuric acid are in the same direction (left), the dissolved zone and oxidized zone are separated, whereas when the fluxes are in the opposite directions they are not separated (right).

sedimentary rocks is caused by the oxidation of included pyrite, except for the slaking of weak sedimentary rocks. Slaking, however, occurs mainly on exposed rock slopes by alternating wetting and drying cycles, which are typically not intense in forested areas. Pyrite oxidation forms sulphuric acid, which in turn dissolves acid-labile minerals, such as calcite and zeolite, and deteriorates rock properties (Fig. 7.5; Chigira and Oyama, 1999). This weathering mechanism forms characteristic weathering profiles based on the directions of groundwater and oxygen fluxes, as shown in Fig. 7.5. The solubility of oxygen dramatically increases with decreasing temperatures (Truesdale *et al.*, 1955), hence snowmelt would contain more oxygen than higher temperature rain. Once oxygen is dissolved in the groundwater at around the freezing point it is captured even though groundwater temperature increases with downward infiltration. Therefore, groundwater recharged from snowmelt can oxidize more pyrite than can groundwater recharged from rain. Thus, climate change could change the weathering rate of rocks. According to a study by Oyama and Chigira (1999), the weathering rate, which is the migration speed of an oxidation front in a vadose zone, can be as high as 80 cm in 80 years. That speed is fast enough to support the idea that present-day weathering affects present-day rock properties and also slope stability.

Reweathering of palaeo weathering profiles and landslides

Weathering profiles usually develop near the ground but are not always a product of the present environment. Some profiles were generated under ancient environments and are now being modified under the present conditions. Such composite weathering profiles are very important for slope stability, because the deeper portions of palaeo weathering profiles could be exposed on slope surfaces by erosion where the rock is weathered again under gravitationally unstable conditions (Chigira and Ito, 1999).

Typical landslides of reweathered materials are observed in granite. Granite is commonly weathered to depths of tens of meters, and many deep weathering profiles are believed to have occurred in geologic time, such as the Tertiary (Ruxton and Berry, 1957; Oyagi, 1968; Kimiya, 1981). Granite is sometimes believed to be very stable and not prone to shallow landslides in previously glaciated regions like northern Europe and northern North America, where weathered materials were mostly eroded away (Durgin, 1977). However, in humid and tropical regions, very thick weathering profiles persist and have been prone to landslides. For example, in Japan, numerous shallow (< a few meters in depth) landslides have occurred on slopes of re-weathered granite, in Kyoto and Shiga in 1953 (Chigira and Ito, 1999), Shimane in 1964 (Oyagi, 1968) and in Aichi in 1972 (Yairi *et al.*, 1972; Kimiya, 1981); these slopes had previously been weathered in the Tertiary or early Pliocene. The slip surfaces of these landslides occurred along the base of extensively loosened granite (Watari and Nakamura, 1970; Iida and Okunishi, 1979; Onda, 1992), although the loosening mechanism has not been elucidated. Many of these landslides occurred following heavy rain on slopes of moderately weathered granite, granite in the strict sense. Durgin (1977) characterized the moderately weathered granite as 'decomposed granite' (not saprolite) and noted that decomposed granite is vulnerable to shallow landslides whereas saprolite is subject to deep rotational slides. According to Chigira and Ito (1999) and Ito (2000), moderately weathered granite is easily weathered to a depth of 50 to 70 cm within 12 years from an artificially cut slope and is characterized by an abrupt weathering front (Fig. 7.6). This re-weathering is due to the disintegration of mineral grains by numerous new micro-cracks. The dominating factors causing this re-weathering are believed to be iterative drying and wetting processes. The reason why such a rapid loosening seldom occurs in heavily weathered granite or granodiorite probably relates to the ability of clayey materials (derived from feldspars, biotite, and hornblend) to support rigid mineral fragments of quartz and potassium feldspar as a matrix. This support prohibits the development of connected micro-crack networks, which are necessary for the loosening. The rate of re-weathering is controlled by climatic conditions, such as rainfall, evaporation, and transpiration, as well as vegetation and rock properties.

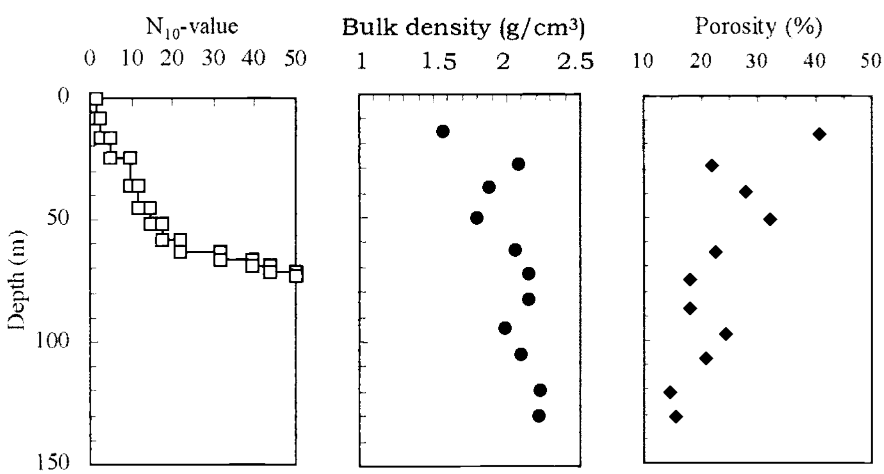


Fig. 7.6. Twelve-year change in portable cone penetration resistance, density and porosity in moderately weathered granite from an artificially cut slope, Shigaraki, Kyoto, Japan (Ito, 2000).

While fracture density and porosity in granite has been shown to increase with degree of weathering, the effect of weathering on permeability appears more complex. Megahan and Clayton (1986) found that saturated hydraulic conductivity was lowest in unweathered granitic material in the Idaho Batholith, USA, due to restricted fracture apertures. Conductivities then increased rapidly in slightly weathered granite. However, as weathering progressed hydraulic conductivities gradually declined, probably due to increasing clay formation and mineral expansion that restricted subsurface flow in both the fractures and rock matrix (Megahan and Clayton, 1986). Such temporal patterns of hydraulic conductivity may significantly influence landslide initiation.

Certain types of granite have clearly recognizable microscopic sheets (micro-sheeting) to a depth of 50 m as the result of unloading, another basic cause of landslides in granite (Fig. 7.7A and 7.7B; Chigira, 2001). A rainstorm in June 1999 generated numerous shallow landslides of weathered granite in Hiroshima Prefecture, Japan (Chigira, 2001). In those landslides, the granite was generally micro-sheeted with low-angle micro-joints dipping gently valleyward. The micro-sheeted granite is easily disintegrated by the interconnection and neoformation of micro-cracks, processes that are accelerated by wetting and drying cycles, temperature change, and creep movements of micro-sheets. This disintegration proceeds with an abrupt front and thus the loosened layer slides (Fig. 7.8, Chigira, 2001). The disintegration rate is not known explicitly, but it is estimated to be in the order of years or tens of years, because micro-sheeted granite is known to deteriorate during the course of construction projects lasting several years.

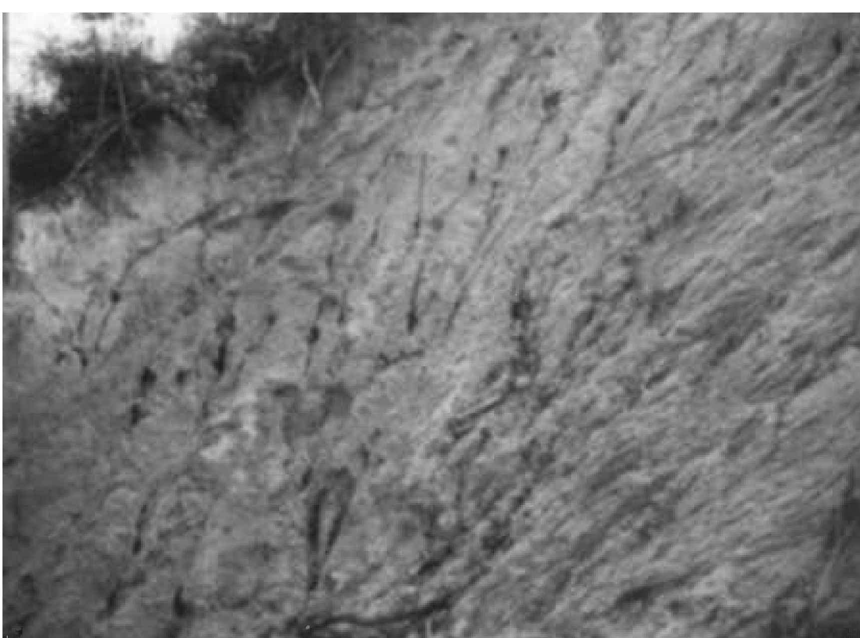


Fig. 7.7A. Landslide of disintegrated granite and debris on micro-sheeted granite (Hiroshima, Japan).

Fig. 7.7B. A close-up view of the micro-sheeted granite (Hiroshima, Japan).

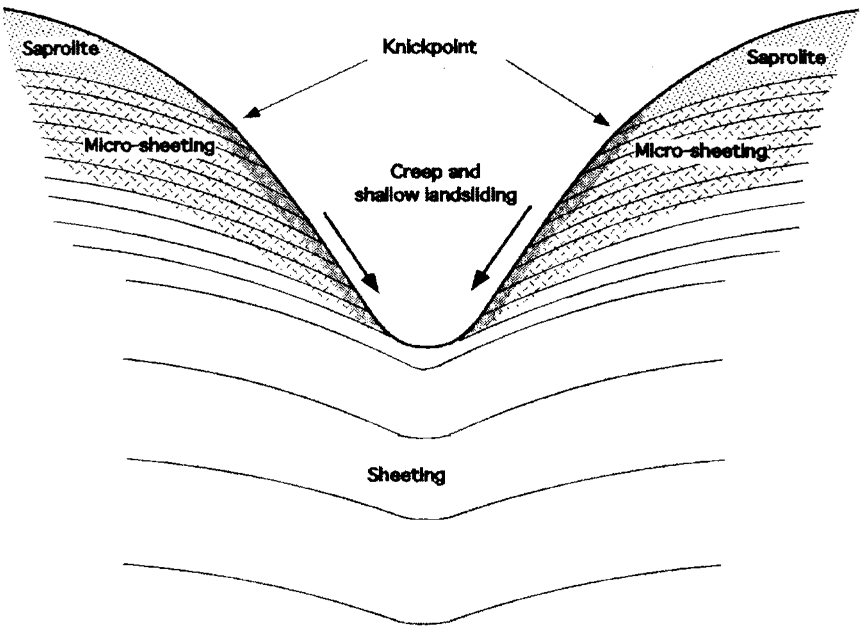


Fig. 7.8. Schematic showing the cross-section of a sheeted and micro-sheeted granite rock slope (Chigira, 2001).

Long-term Gravitational Deformation

Long-term deformation preparing for rapid landslides

Zischinsky (1966) initially studied the gravitational deformation of rock masses and called this deformation 'Sackung'; later Radbruch-Hall (1978) and Chigira (1992) termed this process mass rock creep (MRC). MRC is dominated by flexural slip in the case of foliated rocks, such as sedimentary rocks, slate and schist. Even massive rocks are fractured with new tension or shear fractures by gravitational force and would become rock fragments. Rock masses are thus separated into rock fragments, which are tightly confined in depths of hundreds of meters but are loosened in shallower zones (Fig. 7.9; Chigira, 1995). The deformation sometimes concentrates at a given depth and, in other cases, spreads over various depths beneath a slope. MRC with shearing surfaces at depth leads to essentially no strain in the rock mass near the ground surface. In such cases, environmental changes do not strongly affect the stability of the slope unless groundwater level rises significantly. However, other types of MRC extensively deform rock masses near the ground surface, enabling rock debris to slide. If a flexural slip occurs in a bending fold of steeply dipping, foliated rocks, fractures may develop parallel to or across the foliation (Fig. 7.10; Chigira, 1992). Rock mass is thus heavily fractured and loosened and finally becomes debris. Heavily fractured and loosened rock mass and debris are also created by random tension fractures in massive rocks. Loosened rock mass and rock debris are initially supported by plant roots; however, these would fail during a heavy rainfall or earthquake (Fig. 7.11). Additionally, the rock materials are predisposed to failure when root strength is lost after logging, wildfire, or environmental changes or if MRC proceeds further.

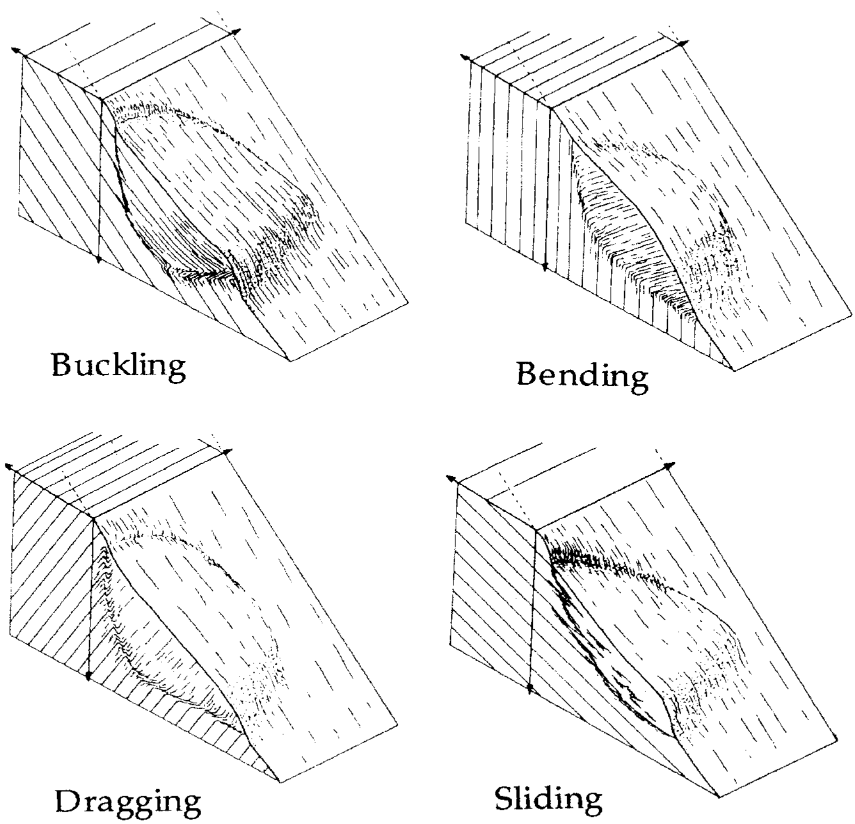


Fig. 7.9. Schematic showing the deformation of mass rock creep in foliated rock slopes (Chigira, 1995).

Fig. 7.10. Toppling type of mass rock creep in slate, Paleogen Setogawa Group, Shizuoka, Japan (Chigira, 1992).

Fig. 7.11. Shallow landslide in slate deformed by mass rock creep, Paleogen Setogawa Group, Shizuoka, Japan (Chigira, 1992).

Besides reacting quickly to environmental changes in the short term as previously described, gravitational deformation of rocks occurs in response to long-term environmental changes. Gravitationally unstable conditions are induced by rapid uplift and erosion. Glacial erosion is one of the most prominent processes; steep slopes supported by the glacier are destabilized when ice melting occurs. Deglaciation can form sheeting joints in valley-wall slopes, and these joints sometimes act as shear surfaces for rock slides (Holzhausen, 1989). Undercutting of slopes by stream erosion or marine wave erosion also generates gravitationally unstable states, sometimes making notches with overhanging rock masses. The rate of notch formation may depend on the intensity of erosion and the durability of the rock.

Landslides caused by changes in groundwater conditions

Climatic change can change the groundwater level, affecting slope stability: a rising groundwater level will destabilize previously stable slopes. There are many cases of palaeo landslides which are inferred to have occurred during climatic conditions different from the present conditions. For example, the Baga Bogd palaeolandslide, in Gobi Altay, Mongolia, is one of the largest landslides in the world, with a volume of 50 km^3 (Phillip and Ritz, 1999). The present climate is dry, but the triggering mechanism for the ancient landslide is believed to be the combination of an earthquake and much higher groundwater levels than presently occur. The Blackhawk landslide, in an arid part of California, USA, is also suspected to have initiated during a wetter period than present (Johnson, 1978). Stable slopes in arid regions could be destabilized by future ascending groundwater. Groundwater levels may change by as much as 200 or 300 m; an argument that has been made in opposition to proposed underground isolation of radioactive wastes at the Yucca Mountain site in Nevada, USA (Anonymous, 1992). Even if subsurface hydrology changes are much smaller, ascending groundwater levels caused by increased precipitation or snowmelt might still be enough to reactivate dormant landslides. Dehn *et al.* (2000) estimated local precipitation by using a climatological model and analyzed the stability of a mudslide in Dolomite, Italy. They estimated that precipitation would decrease in spite of the general greenhouse effect, and concluded that the displacement rate of the mudslide will decrease significantly.

Groundwater levels within slopes along an ocean are affected by changes in sea level rather than changes in precipitation. Such sea-level changes have been occurring in a cyclic manner at intervals of about 100,000 years during the past 700,000 years (Chappel and Shackleton, 1986); currently the sea level is at its highest and is projected to decline. This change, as great as 100 m, occurs in the order of tens of thousands of years, but the Dansgaard-Oeschger cycle occurs in the order of hundreds or thousands of years (Dansgaard *et al.*, 1993).

Even if groundwater level does not change, temperature decreases could freeze rock mass surfaces, leading to rockfall. Freezing of a rock mass surface might form a membrane surrounding the rock mass, preventing the drainage of interstitial water and increasing the bulk weight of the rock mass (Watanabe *et al.*, 1996). A rock mass forming a steep cliff might be thus destabilized. This membrane effect would be stronger for porous rocks, such as sandstone and pyroclastic rocks, than for less porous rocks, because they can contain more water. Many large rockfalls have occurred along the coast of the Japan Sea, and some of them might have been predisposed by such an influence. These rockfalls occurred on steep marine cliffs (higher than 100 m) composed of weak Miocene sandstone and pyroclastic rocks. The Toyohama Tunnel rockfall, with a volume of 11,000 m^3, occurred on 10 February 1996, killing 21 people. The surface part of the cliff was frozen at that time, but groundwater exuded from the detachment surface of the rockfall immediately after its occurrence, suggesting that the groundwater had been dammed up in the fallen rock before the rockfall (Fig. 7.12; Watanabe *et al.*, 1996). Along the detachment surface, old discontinuous joints seem to have existed, and they are believed to be connected to form a single detachment surface. This type of rockfall could occur by the environmental changes accompanying a shift from a temperate to a moderately cold climate. Interstitial water within the rock mass could drain during the temperate climate but could be sealed along the surface of the rock mass. If the weather becomes too cold, the interstitial water within the whole rock body would be frozen and the membrane effect would not occur.

Fig. 7.12. Rockfall at the Toyohama Tunnel, Hokkaido, Japan. Dark areas on the detachment scar are wet areas.

Summary

This chapter summarizes effects of climatic changes on the processes of weathering, slow and long-term gravitational deformation of rocks, and rapid landslides. Weathering, which results in landslide materials, creates weathering profiles during geologic time and sometimes at intervals of tens of years. Additionally, weathering determines the location of potential slip surfaces or landslide behavior. However, the characteristics of weathering profiles are poorly understood at present. Some types of rocks are weathered rapidly and form weathering profiles that favor landsliding. In other words, slopes composed of those rocks are affected by environmental change through weathering. Unwelded ignimbrite is one of the most vulnerable rocks: once its weathered material is stripped by landsliding, weathering commences again to prepare for the next landslide, which can occur in less than a few hundred years. Other types of vulnerable rocks include marine mudstone and sandstone. Granite, weathered granite in a strict sense, is also vulnerable to weathering because weathered granite is re-weathered when it is exposed to a slope surface by erosion. The re-weathering of moderately weathered granite forms an abrupt weathering front, which may comprise a landslide slip surface during heavy rain.

Gravitational deformation of a rock mass (mass rock creep) proceeds gradually but continuously, sometimes concentrating at a specific depth but other times spreading over various depths beneath a slope. The distributed deformation creates many fractures, eventually changing the rock mass to rock debris; the debris is supported by plant roots but may fail if the support is lost after logging, wildfire, or environmental changes, or if MRC proceeds further.

Changes in groundwater conditions affect the stability of slopes. Ascending groundwater decreases slope stability and descending groundwater increases stability. Groundwater that drains from a rock mass under a temperate climate would be locked in if the temperature drops and the resultant freezing of the rock surface creates a membrane surrounding the rock mass. Such a membrane effect could generate a rockfall on a steep rock cliff because of the increase in bulk weight.

References

Anonymous (1992) *Ground water at Yucca Mountain – How high can it rise?* Panel on Coupled Hydrologic/Tectonic/Hydrothermal Systems at Yucca Mountain. National Academy Press, Washington, DC.

Blum, J.D., Gazis, C.A., Jacobson, A.D. and Chamberlain, D.P. (1998) Carbonate versus silicate weathering in the Raikhot watershed within the High Himalayan Crystalline Series. *Geology* 26, 411-414.

Cas, R.A.F. and Wright, J.V. (1996) *Volcanic Successions, Modern and Ancient.* Chapman and Hall, London.

Chappel, J. and Shackleton, N.J. (1986) Oxygen isotopes and sea level. *Nature* 324, 137-140.

Chigira, M. (1990) A mechanism of chemical weathering of mudstone in a mountainous area. *Engineering Geology* 29, 119-138.

Chigira, M. (1992) Long-term gravitational deformation of rocks by mass rock creep. *Engineering Geology* 32, 157-184.

Chigira, M. (1995) *Weathering and Slope Movement*. Kinmiraisha, Nagoya.

Chigira, M. (2001) Micro-sheeting of granite and its relationship with landsliding specifically after the heavy rainstorm in June 1999, Hiroshima Prefecture, Japan. *Engineering Geology*.

Chigira, M. and Inokuchi, T. (in press) Landslides triggered by August 1998 heavy rainfall, northern Japan. *Landslide News*.

Chigira, M. and Ito, E. (1999) Characteristic weathering profiles as basic causes of shallow landslides. In: Yagi, N., Yamagami, T. and Jiang, J.C. (eds), *Slope Stability Engineering*. Vol.2. Balkema, Rotterdam, pp. 1145-1150.

Chigira, M. and Oyama, T. (1999) Mechanism and effect of chemical weathering of soft sedimentary rocks. *Engineering Geology* 55, 3-14.

Chigira, M. and Sone, K. (1991) Chemical weathering mechanisms and their effects on engineering properties of soft sandstone and conglomerate cemented by zeolite in a mountainous area. *Engineering Geology* 30, 195-219

Dansgaard, W., Johnsen, S.J., Clausen, H.B., Dahl-Jensen, D., Gundestrup, N.S., Hammer, C.U., Hividberg, C.S., Steffensen, J.P., Sveinbjornsdottir, A.E., Jounzel, J. and Bond, G. (1993) Evidence for general instability of past climate from 250-kyr ice core record. *Nature* 364, 218-220.

Dearman, W.R. (1974) Weathering classification in the characterisation of rock for engineering purposes in British practice. *Engineering Geology/Geologie de L'Ingenieur* 9, 33-42.

Dehn, M., Burger, G., Buma, J. and Gasparetto, P. (2000) Impact of climate change on slope stability using expanded downscaling. *Engineering Geology* 55, 193-204.

Drever, J.I. (1997) *The Geochemistry of Natural Waters: Surface and Groundwater Environments*. 3rd edition. Prentice-Hall, London.

Durgin, P.B. (1977) Landslides and the weathering of granitic rocks. *Geological Society of America Reviews in Engineering Geology* 3, 127-131.

Folk, R.L. and Patton, E.B. (1982) Buttressed expansion of granite and development of grus in Central Texas. *Zeitschrift fur Geomorphologie* N. F. Bd. 26, 17-32.

Fritz, S.J. (1988) A comparative study of gabbro and granite weathering. *Chemical Geology* 68, 275-290.

Gardner, L.R. (1980) Mobilization of Al and Ti during weathering – isometric geochemical evidence. *Chemical Geology* 30, 151-165.

Holzhausen, G.R. (1989) Origin of sheet structure, 1. Morphology and boundary conditions. *Engineering Geology* 27, 225-278.

Hyman, M.E., Johnson, C.E., Bailey, S.W., April, R.H. and Hornbeck, J.W. (1998) Chemical weathering and cation loss in a base-poor watershed. *Geological Society of America Bulletin* 110, 85-95.

Iida, T. and Okunishi, K. (1979) On the slope development caused by the surface landslides. *Geographical Review, Japan* 52, 426-438.

Johnson, B. (1978) Blackhawk landslide, California, U.S.A. In: Voight, B. (ed.), *Rockslide and Avalanches*. Elsevier, Amsterdam, pp. 481-504.

Ito, E. (2000) *A study on rapid weathering of weathered granite – an example of cut slopes in Shigaraki town, Shiga Prefecture*. Unpublished Master's thesis, Kyoto University.

Kimiya, K. (1981) Distribution of weathering crusts and their formative period in the

Mikawa Highland, central Japan. *Journal of the Geological Society of Japan* 87, 67-69.
Matsukura, Y. (1989) Weathering, rock properties and denudational processes. *Transactions, Japanese Geomorphological Union* 10-A, 103-112.
Megahan, W.F. and Clayton, J.L. (1986) Saturated hydraulic conductivities of granitic materials of the Idaho Batholith. *Journal of Hydrology* 84, 167-180.
Moye, D.G. (1955) Engineering geology for the Snowy Mountain scheme. *Journal of Institution of Engineers, Australia* 27, 287-298.
Nishiyama, K., Yokota, S. and Iwamatsu, A. (1999) Distribution and rock properties of reddish weathered gravels within the Quaternary fan deposits. *Journal of the Japan Society of Engineering Geology* 40, 2-13.
Onda, Y. (1992) Influence of water storage capacity in the regolith zone on hydrological characteristics, slope processes, and slope form. *Zeitschrift für Geomorphologie N.F.* 36, 165-178.
Oyagi, N. (1968) Weathering-zone structure and landslides of the area of granitic rocks in Kamo-Daito, Shimane Prefecture. *Reports of cooperative research for disaster prevention by National Research Center for Disaster Prevention* 14, 113-127.
Oyama, T. and Chigira, M. (1999) Weathering rate of mudstone and tuff on old unlined tunnel walls. *Engineering Geology* 55, 15-28.
Phillip, H. and Ritz, J.-F. (1999) Gigantic paleolandslide associated with active faulting along the Bogd fault (Gobi-Altay, Mongolia). *Geology* 27, 211-214.
Petit, J.C., Mea, G.D., Dran, J.C., Magonthier, M.C., Mando, P.A. and Paccagella, A. (1990) Hydrated-layer formation during dissolution of complex silicate glasses and minerals. *Geochimica Cosmochimica Acta* 54, 1941-1955.
Pye, K. and Miller, J.A. (1990) Chemical and biochemical weathering of pyritic mudrocks in a shale embankment. *Quarterly Journal of Engineering Geology, London* 23, 365-381.
Radbruch-Hall, D.H. (1978) Gravitational creep of rock masses on slopes. In: Voight, B. (ed.), *Rockslides and Avalanches, Vol. 1 Natural Phenomena*. Elsevier, Amsterdam, pp. 607-657.
Raymo, M.E. and Ruddiman, W.F. (1992) Tectonic forcing of late Cenozoic climate. *Nature* 359, 117-122.
Ruxton, B.P. and Berry, L. (1957) Weathering of granite and associated erosional features in Hong Kong. *Bulletin of the Geological Society of America* 68, 1263-1292.
Schrott, L. and Pasuto, A. (eds) (1999) Temporal stability and activity of landslides in Europe with respect to climatic change (TELSEC). *Geomorphology* 30, 1-211.
Selby, M.J. (1993) *Hillslope Materials and Processes*. Oxford University Press, Oxford, UK.
Shimokawa, E., Jitousono, T. and Takano, S. (1989) Periodicity of shallow landslide on Shirasu (Ito pyroclastic flow deposits) steep slopes and prediction of potential landslide sites. *Transactions, Japanese Geomorphological Union* 10, 267-284.
Suzuki, T. (1992) Tephrochronological study on Nasu volcano. *Bulletin of the Volcanological Society of Japan* 37, 251-263.
Truesdale, G.A., Downing, A.L. and Lowden, G.F. (1955) The solubility of oxygen in pure water and in sea water. *Journal of Applied Chemistry* 5, 53-62.
Watanabe, T., Minoura, N., Ui, T., Kawamura, M., Fujiwara, Y. and Matsueda, H. (1996) Geology of a collapse of the sea-cliff at the western entrance of the Toyohama Tunnel, Hokkaido, Japan. *Journal of Natural Disaster Science* 18, 73-87.
Watari, M. and Nakamura, H. (1970) Stability of steep slopes on weathered granite in

Kure and Kobe Cities. *Reports of cooperative research for disaster prevention by National Research Center for Disaster Prevention* 24, 127-1369.
White, A.F. and Brantley, S.L. (eds) (1995) Chemical weathering rates of silicate minerals. *Reviews in Mineralogy*, Mineralogical Society of America, vol. 31.
White, A.F., Blum, A.E., Schulz, M.S., Vivit, D.V., Stonestrom, D.A., Larsen, U.S., Murphy, S.F. and Eberl, D. (1998) Chemical weathering in a tropical watershed, luquillo mountains, Puerto Rico: I. Long-term versus short-term weathering fluxes. *Geochimica Cosmochimica Acta* 62, 209-226.
Winkler, E.M. (1994) *Stone in architecture – Properties, Durability*. Springer-Verlag, Germany.
Yairi, K., Suwa, K. and Masuoka, Y. (1972) Landslides by the heavy rainfall of July 1972 – Disaster in Obara and Fujioka Villages, Aichi Prefecture. In: K. Yabe (ed.), *Report of the Grant-in Aid for Scientific Research from the Japanese Ministry of Education, Science, Culture and Sports* pp. 92-103.
Yatsu, E. (1988) *The Nature of Weathering - an Introduction*. Sozosha, Tokyo.
Yokota, S. (1999) Weathering distribution in a steep slope of soft pyroclastic rocks as an indicator of slope instability. *Engineering Geology* 55, 57-68.
Zischinsky, U. (1966) On the deformation of high slopes. In: *Proceedings of the First Congress of the International Society of Rock Mechanics, Lisbon, Portugal*. Vol. 2, pp. 179-185.

Chapter 8

Potential Effects of Environmental Change on Landslide Hazards in Forest Environments

Roy C. Sidle[1] and Amod S. Dhakal[2]
[1]*Professor, Department of Geography, The National University of Singapore, Republic of Singapore*
[2]*Killam Postdoctoral Research Fellow, Department of Forest Resources Management, University of British Columbia, Vancouver, Canada*

The interactive roles of the natural factors that contribute to landslide initiation and potential climate changes are complex. Additional complexity is added when changing forest land uses are considered. Thus, while it is impossible to ascertain specific patterns of landslide response to possible climate change, certain generalizations can be made and scenarios can be discussed. Shallow, rapid landslides will only increase in scenarios of increasing rain event intensity, whereas, deep-seated mass movements will increase with seasonal increases in precipitation. Dry ravel may respond more directly to warming, increasing with sparse vegetation covers and increased frequency of fire. Climate change has the greatest impacts on landslide occurrence by modifying evapotranspiration and root strength of vegetation. Evapotranspiration affects soil water recharge and subsurface flow and thus influences slope stability. Climate change may alter canopy structure, soil runoff and evaporation, ground cover, and rooting depth; these in turn affect water storage and routing in unstable sites. Anthropogenic and climate-induced changes in vegetation cover particularly affect the potential for shallow landsliding due to modifications in root cohesion. While much attention has been placed on the impacts of forest harvesting practices on increased landslide erosion, the effects of permanent forest land conversion, particularly in the tropics, appear more problematic. Additionally, the increasing impacts of affluent recreation in the developing world need to be assessed. Agroforestry may offer some socio-economic and natural resource

advantages in selected cases, but cumulative off-site effects must be evaluated. Depending on the level of detail required and type of available data, landslide hazard can be assessed by several methods: (i) terrain stability classification; (ii) empirical landslide hazard assessment; and (iii) physically-based models. Including climatic change scenarios into landslide hazard models is a complex task because scenarios of climatic change are unclear even at regional levels. Also, most models only evaluate land use in a very general way. Thus, much progress is needed to predict these interactive climate-change and anthropogenic impacts on slope stability.

Introduction

Landslides pose significant hazards to humans and property (e.g., Li, 1989; Swanston and Schuster, 1989; Yoon, 1991) as well as contributing substantial supplies of sediment to receiving streams and river systems (e.g., Pearce and Watson, 1983; Lamberti *et al.*, 1991; Reneau and Dietrich, 1991). Of all the natural hazards that affect forested areas, landslides are among the most common and damaging. Regions with high precipitation, steep hillslopes, tectonic activity, fragile geology, and widespread land management and development are prone to landslide hazards (Eisbacher and Clague, 1984; Sidle *et al.*, 1985). While significant on-site disturbance results from widespread mass movement (including loss in site productivity; see Fig. 8.1A), human casualties and property damage (Figs 8.1B and 8.1D), as well as deterioration of channels and aquatic habitat (Fig. 8.1C) typically occur in depositional and scour zones of large debris flows triggered by landslides.

Aside from other natural hazards such as earthquakes and volcanic eruptions that can trigger landslides, most slope failures in forested terrain initiate or activate when water recharges the soil mantle or regolith, either during excessive rainfall or snowmelt or a combination of the two (Sidle and Swanston, 1982; Tsukamoto *et al.*, 1982; Megahan, 1983; Aniya, 1985; Wieczorek *et al.*, 1989; Bangxing *et al.*, 1994). In steep terrain, the delicate balance between resistance of the shallow soil mantle to failure and the gravitational forces tending to move the soil downslope can easily be upset by rapid inputs of water. In addition to the primary influence of pore pressure on shallow landslide initiation, vegetation condition, soil properties, and geomorphology affect site stability.

The physical and biological components of terrestrial ecosystems are very sensitive to climate because climate determines the distribution of plants and formation of soils via weathering of geological materials. Changes in the world's climate due to increases in concentrations of carbon dioxide (CO_2) and other atmospheric gases (e.g., methane, nitrous oxide) can potentially affect landslide initiation throughout the world in many ways. The greenhouse phenomenon and the presence of greenhouse gases in the atmosphere are vital controls on the temperature of the earth. If these gases were to increase, temperatures would rise; decreases in these gases would cause temperatures to fall. There is a consensus on the relative responses but not the magnitudes of surface warming and stratospheric cooling related to greenhouse gases (Lindzen, 1990).

Fig. 8.1. Landslides and resultant environmental and human impacts. Different impact forms of landslides resulting from a heavy rainfall in the central Nepal (Palung area) are shown. The area received 340 mm of rain with a maximum hourly rainfall of 65 mm on 19 July 1993. A – Shallow landslides on the vegetated slopes; B – Cobble-type debris flow deposition zone. The debris flow was triggered by numerous landslides on the upper slopes and claimed 58 lives in this depositional area; C – Sand deposition area caused by flooding and the debris flow from the upper catchment; D – Boulder-type debris flow caused by landslide initiation and channel scouring that claimed 10 lives.

The most common global climate change scenarios associated with geomorphic hazards are increasing mean air temperature and changes in regional annual and seasonal precipitation (Zimmermann and Haeberli, 1992; Sidorova, 1998; Buma, 2000). Although certain types of landslides may be influenced by these regional climate changes, the initiation and persistence of slope failures are more related to the timing and short-term perturbations of rainfall and snowmelt.

Landslides could respond rather quickly to new climate conditions even though our knowledge about how current climate conditions relate to geomorphic process is rather limited (Boer *et al.*, 1990; Schlyter *et al.*, 1993;

Evans and Clague, 1994; Sidorova, 1998; Dehn and Buma, 1999; Buma, 2000). Long-term changes in average climate conditions (temperature and precipitation) as well as possible shifts in the frequency of extreme events are expected as a result of climate change. These factors will play an important role in influencing landslide scenarios throughout the world (Evans and Clauge, 1994; Wyss and Yim, 1996). The effect of climate change on other environmental factors, such as vegetation and soil, may introduce more complex interactions and scenarios related to landslide occurrence.

This chapter provides a broad overview of climate change as it relates to various types of landslide hazards in forested terrain. Insights into how forest practices and forest conversion influence slope stability are discussed along with how management can assess landslide hazards and adapt to challenges posed by possible future climate and related environmental changes. Examples of two landslide hazard assessment methods suitable at different scales of analysis are presented.

Future Climate and Environmental Change Scenarios

Evidence and models

There is strong evidence that the levels of several atmospheric greenhouse gases have increased during the past 150 years (Ciesla, 1995). Current climate models predict a global warming of about 2°C in the 110-year period from 1990 to 2100 (Acosta *et al.*, 1999). This projection accounts for the effects of aerosols and the influence of oceanic inertia where the earth's surface and lower atmosphere would continue to warm by 1-2°C even if greenhouse gas concentrations stop rising in 2100.

General circulation models (GCMs) have been widely used to simulate global climatic sensitivity to increased greenhouse gases (Cess *et al.*, 1990). GCMs are physically based and use the laws of conservation of mass (related to water vapour and air), momentum, and heat in the atmosphere, together with state equations, which are then solved for atmospheric variables such as wind speed and direction, temperature, humidity, surface pressure, and precipitation (Loaiciga *et al.*, 1996). A major limitation of these coupled models for predicting geomorphic hazards is that deep oceanic mixing may occur on timeframes of up to 1000 years while atmospheric responses of interest to landslide initiation may occur in fractions of an hour. Additionally, limitations related to spatial resolution of GCMs are significant in terms of landslide hazard analysis (Buma and Dehn, 1998). Recent efforts to extract hydrologic information useful at the river basin scale from GCMs using various nested schemes partially address these spatial scaling limitations (e.g., Wood *et al.*, 1992; Vorosmarty *et al.*, 1993). Furthermore, incomplete understanding of the roles of soils and especially vegetation feedback related to climatic warming makes predictions of local and regional forest harvesting scenarios virtually impossible (Manabe and Wetherald, 1987; Loaiciga *et al.*, 1996). As such, it appears that GCMs provide only background information and general climatic trend data that may be useful in framing the comparisons or scenarios of landslide response in forest environments subject to long-term climatic change.

Temperature changes attributed to greenhouse gases have been found to induce changes in precipitation, number of frost-free days, and the frequency and severity of storms (Bello, 1998; Francis and Hengeveld, 1998; Naranjo-Diaj and Centella, 1998). Since vegetation utilizes CO_2 during photosynthesis, increased levels of CO_2 have potentially significant impacts on the growth, survival, and adaptations of forest communities. In addition, changes in climate could affect the processes involved in soil formation. These varying climatic attributes that influence landslide initiation are discussed in the following paragraphs.

Storm magnitudes

The frequency and intensity of extreme weather events such as storms and hurricanes are expected to change with scenarios of climatic warming. Already there is scattered evidence of recent patterns of increases in extreme precipitation events (Mason *et al.*, 1999; Zhai *et al.*, 1999; Pfister *et al.*, 2000), however, the linkage of such increases to climate change is clouded by the lack of widespread long-term precipitation records, variable data quality, decadal variability, and spatially offsetting effects during El Niño events (Easterling *et al.*, 1999; Fu and Wen, 1999; Landsea *et al.*, 1999; Trenberth and Owen, 1999). All regions of the world have experienced extreme climatic events from time to time. Many such events were observed during the last century in different parts of the world. During the 1990s alone, at least half a dozen floods of epic proportions have occurred in major drainage basins of Canada and the USA, central Europe, and southern China (Francis and Hengeveld, 1998).

It has been suggested that Europe will experience a higher frequency of extreme rainfall events in future years (Rowntree, 1993; Cubasch *et al.*, 1995). For such scenarios, rainfall tends to be more intense with greater increases in rainfall on the wettest days of the season compared to smaller increases in mean rainfall. These increases are presumably due to the higher water content of warmer air (Rowntree, 1993). The year-to-year variability (standard deviation of seasonal precipitation) also generally increases more than the mean. Although these findings were based on an analysis confined to mid-latitude regions of Europe, an analysis north of 45°N yields similar results with findings indicating a possible increase in landsliding in that region. The Intergovernmental Panel on Climate Change (IPCC) initially estimated that tropical and subtropical regions would experience the least alterations in precipitation regime (IPCC, 1990); however, more recent scenarios developed for the greater tropics indicate a possibility for spatially explicit seasonal increases (from tropical north Africa into India) and decreases (southern and western Africa, central and western Australia and the Amazon Basin) in precipitation (Hulme and Viner, 1998). However, for many regions, models still cannot predict how extreme precipitation events will respond to climate change. Since models used to simulate such climatic changes cannot themselves simulate extreme weather events, the evidence outlined here is indirect and inferential.

Annual and seasonal precipitation regimes

Winter precipitation in regions north of 45°N is expected to rise (Mansell, 1997). Simulations of river basins by means of macroscale hydrologic models nested within GCMs in mid-latitudinal basins of the USA predict shorter winter seasons, larger winter floods, drier and longer summers, and overall enhanced and protracted hydrologic variability (Loaiciga *et al.*, 1996). While most of these predictions would exacerbate landslide potential, the seasonal period of rainfall-induced landslide susceptibility may decrease in mid-latitude regions. In subtropical latitudes, annual precipitation changes are estimated to be minimal (IPCC, 1990), although at least some scenarios of precipitation change have been simulated (Hulme and Viner, 1998). Though the prediction depends on the details of the particular climate model and the emissions scenario, the effect of aerosols may significantly weaken the Asian summer monsoon (Acosta *et al.*, 1999). Thus, these scenarios of annual and seasonal rainfall response to climate change are not expected to have a large influence on landslide initiation in the tropics and subtropics. Global average precipitation is estimated to increase in the range of 3-15% from current levels (IPCC, 1990). Although both globally-averaged total annual precipitation and evapotranspiration are predicted to increase, local trends are much less certain (Mimikou *et al.*, 1991). Climate models are still unable to precisely predict regional patterns of annual precipitation (Loaiciga *et al.*, 1996). In addition, the hydrological cycle is extremely complex: a change in precipitation may affect surface wetness, reflectivity, and vegetation, which then influences evapotranspiration, which in turn affects precipitation.

Snow accumulation and melt

Winter snowpacks are projected to decline in most regions south of 60°N (Rowntree, 1993). Portions of Scandinavia and southern Alaska could also have reduced winter snowpacks (Rowntree, 1993). Arctic areas would have increased winter snow accumulation in most climate-change scenarios (Rowntree, 1993). Increased water inputs to the soil as affected by snow depth and melt timing could accelerate solufluction processes (Rowntree, 1993). Otherwise, this projected increase in Arctic snow mass associated with the general increase in precipitation has a greater influence on the snow budget. Faster melting of the Greenland and Antarctica ice sheets is likely to be balanced by increased snowfall in both regions.

The snowmelt dynamics could influence landslide initiation in the regions south of 60°N. Because snow accumulation is less, snowpacks may ripen and melt earlier with less meltwater released. Consequently, landslides triggered by snowmelt may decline. However, the timing of snowmelt related to the thermal conditioning of the snowpack is critical to episodic melt rates that trigger landslides. Such short-term dynamics (including rain-on-snow events) cannot be estimated in current climate-change forecasts. Alternatively, increases in debris flow activity within periglacial areas (particularly at the lower boundary) can occur as the result of the disappearance of perennial snow patches. Zimmermann *et al.* (1986) report such increases in the Swiss Alps. The influence of glacial shrinkages and degradation of permafrost on soil mass movement is discussed in detail in another chapter of this volume (Haeberli and Burn, Chapter 9).

Soil moisture conditions

More rain and snow will cause wetter soil conditions in the high-latitudes during winter, but higher temperatures may mean drier soils in the summer (Acosta *et al.*, 1999). Increases in soil moisture should generally be less widespread than increases in precipitation because evaporation is predicted to increase from spring onwards due to earlier snowmelt and higher temperatures (Rowntree, 1993). In general, climatic models predict that both evaporation and precipitation will increase in most regions. While some regions may become wetter, in others the net effect of the temperature-modified hydrological cycle will be a loss of soil moisture. Climate-change simulations in central Greece indicate that soil moisture during winter will remain unchanged for all climatic scenarios, whereas significant soil moisture decreases may occur in summer (Panagoulia and Dimou, 1997). The magnitude of these results varies somewhat depending on the climatic model employed. Other climatic models suggest that convective storms will become more intense. This condition, coupled with seasonal climatic drying, would lead to increased overland flow, especially in areas where the infiltration capacity is low to moderate, thus increasing flood and surface erosion hazards. Although local changes in soil moisture are important for slope stability, the spatial resolution of existing distributed soil moisture models severely limits any application related to landslide hazard assessment. The use of topography (e.g., geomorphic hollows) as a surrogate for spatially explicit soil moisture (e.g., Anderson and Burt, 1977; Tsuboyama *et al.*, 1994, 1998) appears to provide better spatial resolution of this important parameter; however, such details cannot be incorporated into climate change scenarios.

Studies to date suggest that only minimal changes in soil moisture conditions due to climate change would be expected in temperate regions with winter rainstorms. Thus, impacts on slope stability would be negligible. On the other hand, regions that experience landslides during summer convective storms may have drier soils and therefore would be less susceptible to landslides. Conversely, if storm intensity increases, as has been suggested in some regions (Loaiciga *et al.*, 1996), this may more than compensate for any drier soil conditions attributed to climatic warming. Any realized reduction in landslide hazard would be most evident during the first large rainstorm of the season (i.e., driest soil conditions) (Sidle *et al.*, 1985).

Vegetation changes

A warming of 1-3.5°C during the next 100 years would shift current climate zones poleward by approximately 150 to 550 km and vertically by 150-550 m in mid-latitude regions (Acosta *et al.*, 1999). Observations, experiments, and models demonstrate that a sustained increase of just 1°C in the global average temperature would affect the functioning and composition of forests. A typical climate change scenario for the 21st century indicates a major impact on the species composition of one third of world's existing forest (varying by region from 14 to 67%; Acosta *et al.*, 1999). Entire forest types may disappear, while new combinations of species and hence new ecosystems, may be established. Since higher latitudes are expected to warm more than equatorial zones, boreal forests will be more affected than temperate and tropical forests (Acosta *et al.*, 1999). Vegetation species and ecosystems are forced to migrate to higher elevations in response to predicted climate change. However, empirical evidence from western North America suggests that changes in tree lines may be very subtle (Peterson, 1998) while, in other cases, changes in the distribution of forests will be affected by the availability of suitable sites for tree growth (Beniston and Innes, 1998).

Because vegetation rooting strength is highly species dependent (Sidle, 1991), the influence of forest-cover changes on landslide susceptibility depends on species composition as well as stand density and biomass of the evolving forest. Generally, deciduous species have stronger rooting systems than conifers, however, attributes such as below-ground biomass, percentage of larger roots, depth of root penetration, and spacing of individual stems all influence the net root-strength contribution (Wu *et al.*, 1979; Gray and Megahan, 1981; Sidle, 1992).

The productivity and growing seasons of rangelands and pastures would be affected by global warming. Shifts in temperature and precipitation may reshape the boundaries between grasslands, shrub lands, forests and other ecosystems. In temperate and subtropical grasslands, an extended growing season together with increased evapotranspiration may actually reduce landslide potential (or at least the seasonal period of greatest landslide susceptibility). These benefits may be realized in areas such as North Island, New Zealand, where widespread conversion of native forests to pastureland occurred more than a century ago. Of course, such sites would remain relatively unstable compared to previous forest cover (Selby, 1974). In contrast, tropical grasslands may be transformed to desert or temperate grasslands and thus reduce the net rooting strength in the soil mantle. While rooting strength of grasses is typically very small compared to trees and shrubs (Sidle *et al.*, 1985), transformation to a more sparse grass cover may lend former tropical ecosystems more prone to slope failure. Furthermore, the potential migration of grasslands to higher elevations formerly occupied by forests and shrubs, may increase landslide potential due to the introduction of weaker rooted species on steep uplands. An analogy is the practice of slash and burn agriculture in the tropics of Asia and Latin America; site stability is reduced when steep forestlands are converted to temporary cropland with weak root strength characteristics (Wright and Mella, 1963; Starkel, 1972).

Land use practices and demographics

Vegetation response, survival, and adaptation will be variable in climate warming scenarios. Plant growth and health may benefit from fewer freezing periods, but damage can occur in certain types of vegetation at higher temperatures, particularly when combined with water stresses (Acosta *et al.*, 1999). Certain weeds may expand their range into higher-latitude habitats. There is also some evidence that the poleward expansion of insects and plant diseases will increase the risk of losing certain vegetation types (Acosta *et al.*, 1999). Future changes in climate will lead to a spatial shift of crop productivity potential as well as forest types. Such shifts may require relocation of cropping patterns and forest management strategies from the present situations.

The people most vulnerable to climate change are the landless, poor, and isolated. Poor mechanisms of trade, weak infrastructure, lack of access to technology and information, and armed conflict will make it more difficult for poor people to adapt land use practices to changing climates. Many of the world's poorest isolated areas face the greatest risk: sub-Saharan Africa; South, East and Southeast Asia; tropical areas of Latin America; and some Pacific Island nations. Decline in productivity of natural resources in rural areas may accelerate rural-to-urban migration in the developing world. The resultant declines in productivity could increase the probability of landsliding due to reduced vegetation rooting strength (Temple and Rapp, 1972). In other cases, rural farmers may relocate to higher elevation sites in response to climate change, thus exacerbating the landslide hazard (Lanly, 1969; Eckholm, 1979).

Climate Thresholds for Different Types of Mass Soil Movements

Categorization of mass movements

Slope movements have been classified in many ways, each having some usefulness related to the recognition, avoidance, control, or correction of the hazard. The widely used classification scheme developed by Varnes (1978) distinguishes five types of mass movement (falls, topples, slides, spreads, and flows) plus combinations of these principal types along with the type of material (bedrock and engineering soil). In this chapter we simplify this scheme and broadly classify mass movements on forested hillslopes into three major functional groups: (i) shallow, rapid failures (debris slides, avalanches, and flows); (ii) slow, generally deeper mass movements (slumps, earthflows, and soil creep); and (iii) surficial mass wasting (dry ravel, dry creep).

Table 8.1. Climatic factors that initiate and activate different types of mass soil movements and likely effects of two simple climate change scenarios on such mass movements.

	Shallow, rapid landslides	Slow, deep-seated landslides	Surficial mass soil movement
Climatic triggering factors	High intensity rainstorms Wet antecedent conditions Rapid and abundant snowmelt	Prolonged (seasonal) accretion of soil moisture Prolonged snowmelt	Increased freeze-thaw wetting-drying cycles Increased temperatures promote fire
Climate Change Scenarios			
More extreme rain events, but little change in total precipitation	Increase in frequency and magnitude	Little effect	May increase in conjunction with a storm just after fire
More total rain and snow, but higher temperatures	Little effect for rainfall-triggered landslides, but may expand the window of susceptibility; snowmelt triggered failures may increase and occur earlier	The rate of movement for deep-seated failures will likely increase; the period of movement could expand past the rainy season but could initiate later if summer antecedent moisture is low; period of movement in areas with snowpacks may decrease	Where temperature increases promote fire or increased freezing and thawing, dry ravel and dry creep may increase; under warmer conditions, less wetting and drying may sometimes occur, thus ravel would decline
Less snow accumulation at mid-latitudes	Decrease in frequency, but occur earlier in the melt season	Decrease the period and magnitude of movement	Little effect

Climatic characteristics, which play the dominant role in triggering these three kinds of failures, are usually different. Additionally, thresholds of landslide response to climatic inputs differ across failure types (Table 8.1). Thus, landslide hazard assessment procedures require different climatic and hydrologic inputs for various failure types; inputs may range from yearly, monthly or daily precipitation to short-term intensities in intervals of hours or minutes. Hence, our broad categorization of soil mass movements is useful to assess the differential impacts of climate change.

Shallow, rapid failures

Fig. 8.2. Debris avalanches and subsequent debris flows resulting from severe typhoon storms in August and September 1993 in the Kagoshima Bay area, Kyushu, Japan. These disasters caused 119 fatalities (photo by R.C. Sidle).

Debris slides, debris avalanches, and debris flows are typical shallow movement types in steep forested landscapes. In many sites, these translational failures initiate as a slower moving debris slide. Water may then be incorporated into the failure mass transforming it into a more rapidly moving debris avalanche (0.3 m min^{-1} to > 3 m s^{-1}). With increasing liquefaction and channelization, the failure may then become a very rapid (> 3 m s^{-1}) debris flow (Temple and Rapp, 1972; Sidle *et al.*, 1985) (Fig. 8.2). During major rainstorms or snowmelt events, a pore water pressure builds up in the soil mantle usually just above the lithic contact or other hydrologic impeding layer (e.g., Onda *et al.*, 1992). A common hydrologic sequence for shallow landsliding involves wet antecedent conditions followed by a prolonged period of rainfall with a burst of high intensity (e.g., Sidle and Swanston, 1982) (Table 8.1). Hence, these failures are largely dependent on the characteristics of individual storms, particularly intensity and, to a lesser extent, total storm precipitation (Sidle, 1986). Additionally, the extent of pore pressure response is influenced by antecedent moisture (Sidle and Tsuboyama, 1992; Tsuboyama *et al.*, 2000). Thus, in regions of the world where the frequency and intensity of extreme rainfall are expected to increase with changes in climate, greater numbers of shallow, rapid landslides are likely.

Slow, deep-seated mass movements

Fig. 8.3. A large combination slump-earthflow in the Oregon Coast Ranges, USA (photo by R.C. Sidle).

Both active and dormant deep-seated mass movements characteristically occur in somewhat gently sloping topography that is typically hummocky with immature drainage systems (Sidle *et al.*, 1985). Failure or deformation occurs in deep, heavily weathered clay-rich soils or regoliths (Kelsey *et al.*, 1981). Slumps and earthflows often occur in combination; the initial movement is typically a slump, and subsequent downslope movement of the remoulded material is by earthflow (Fig. 8.3). Soil creep is the plastic deformation of the soil mantle and, thus, no failure plane is present (Sidle *et al.*, 1985). Creep actually occurs in all hillslope soil mantles regardless of depth, but it is most significant where soils are clay-rich and deeply weathered. Movement rates of deep-seated failures range from millimetres per year (extremely slow) to metres per day (rapid), although slower rates are more typical (Swanson and Swanston, 1977; Wasson and Hall, 1982; Sidle *et al.*, 1985). Despite their slower rates, such slope movements are responsible for the transport of large volumes of sediment to streams and rivers in certain regions (Gregory and Walling, 1973; Dietrich and Dunne, 1978).

Deep-seated mass movements are more influenced by seasonal or annual variations in water inputs due to the deeply weathered nature of the regolith (Campbell, 1966; Swanson and Swanston, 1977; Bechini, 1993). Typically, deep-seated movements respond non-linearly to accumulations of soil water; during the dry season little if any movement occurs in response to rainfall, but once the regolith is recharged to a critical point, mass movement responds to further rainfall or snowmelt inputs (Oyagi, 1977; Swanson and Swanston, 1977) (Table 8.1). With predicted increases in winter precipitation in northern latitude regions (above 45° N), the rate of movement in some winter rainfall and rain-on-snow dominated areas may increase. However, with predicted shorter winters, the actual period of deep-seated mass movement may decrease. Increased temperatures may accelerate weathering processes and tend to increase the susceptibility and rate of deep-seated mass movement. The amount and timing of water released from spring snowmelt will affect deep-seated mass movements. In mid-latitude regions where the snowpack is expected to decrease, deep-seated mass movements will occur earlier but for a shorter period of time and probably at a reduced seasonal rate of movement. In northern latitudes where snowpacks increase, mass movements will probably occur earlier (due to climate warming), have increased seasonal rates of movement, and have a longer period of activity (but not necessarily further into the summer). Bovis and Jones (1992) used recent climate change records, dendrochronological data, and stratigraphic records to show that movement of large earthflows in British Columbia, Canada, responded to Holocene hydroclimatic changes.

Surficial mass wasting

Surficial mass wasting (dry ravel and dry creep) involves the downslope movement by gravity of individual soil grains, aggregates, and coarse fragments (Fig. 8.4). The particles generally move by rolling, sliding, or bounding down steep slopes (Sidle *et al.*, 1985). The main cause of dry ravel is commonly believed to be the loss of interlocking frictional resistance among soil aggregates or grains (e.g., Hough, 1951; Rahn, 1969), which may occur in relation to freezing-thawing (Asare *et al.*, 1997) and wetting-drying cycles; however, recent research suggests that slightly deeper sliding failures in granular materials may be triggered by the successive occurrence of rolling of near-surface particles (Onda and Matsukura, 1997).

Fig. 8.4. Dry ravel occurring after fire on a steep harvested slope that was recently burned. Note that ravel is trapped on the upslope side of stumps (photo by R.C. Sidle).

While surficial mass wasting processes transport far less sediment than other landslide types, they can pose problems in areas with sparse vegetative cover, especially where soils have been disturbed (Megahan, 1978; Sidle *et al.*, 1993). In particular, fire can induce extensive ravelling in certain landscapes (Krammes, 1965; Mersereau and Dyrness, 1972; Rice, 1982; Florsheim *et al.*, 1991) (Fig. 8.4). Thus, any climate change scenario in steep terrain that exacerbates wildfire occurrence would probably increase the rates of dry ravel and dry creep. Since higher temperatures due to climatic change are likely to affect high elevation permafrost distribution, thawing of permafrost may also induce ravelling and dry creep at higher elevation sites. Any climate changes that promote more frequent freezing-thawing or wetting-drying cycles would induce greater dry ravel and dry creep on steep disturbed or partially vegetated hillslopes (Table 8.1). Conversely, fewer of these cycles would tend to reduce dry ravel and dry creep.

Effects of Changes in Vegetation

Vegetation response to increases in atmospheric CO_2 and to resulting increases in surface temperature can be assessed at scales ranging from the microscopic cellular level to the macroscopic ecosystem level (Rosenzweig and Hillel, 1998).

The processes of photosynthesis, respiration, and transpiration within plants are most directly affected by climate change. The most significant direct and indirect influences of these physiological changes on landslide initiation relate to evapotranspiration and root strength.

Evapotranspiration

Evapotranspiration influences soil water recharge and subsurface flow and thus has the potential to affect slope stability. Three factors comprise evapotranspiration from vegetated landscapes: (i) evaporation from the soil surface; (ii) transpiration; and (iii) interception and subsequent evaporation from canopies or foliage. Evaporation from the soil surface is controlled by the depth to water table, soil pore structure, and local heat budgets (Jones, 1997). For densely vegetated landscapes, variations in the amount of evaporation directly from the soil surface are small compared with variations in other evapotranspiration components (Sidle *et al.*, 1985). Although difficult to assess in experiments, evaporation from soil surfaces under well-vegetated canopies may comprise approximately 10% or slightly more of total evapotranspiration; proportional rates in dryland areas with sparse vegetation are much higher (Campbell, 1985). Transpiration is the dominant process in the conversion of forest soil moisture to water vapour (Reifsnyder and Lull, 1965). It involves the adsorption of soil water by plant roots, subsequent translocation of the water through the plant, and release of water in the vapour phase through stomatal openings in the foliage. Transpiration rates in vegetated landscapes depend on availability of solar energy and soil moisture as well as vegetation characteristics, including stand density, height and age, albedo of the foliage, and canopy structure (Reifsnyder and Lull, 1965). Nevertheless, rates of transpiration are rather similar for different vegetation types if water is freely available (e.g., Monteith, 1976). Interception and subsequent evaporation of water from vegetation cover is particularly significant in coniferous forests; losses (both snow and rain) from these dense canopies can account for up to 30-50% of gross annual precipitation (Dingman, 1994). Seasonal dormancy of deciduous species will greatly reduce canopy interception, often when the heaviest precipitation occurs; however, multi-tiered tropical forest canopies have a rather high potential to intercept water due to their structure and evergreen nature. Since it is difficult to measure the separate fluxes of soil surface evaporation, transpiration, and interception, these losses are typically lumped together as evapotranspiration. In general, evapotranspiration rates in temperate regions are lowest from bare soil, several times higher from grassland, and 5 to 10 times higher (compared to bare soil) from forests (Jones, 1997).

The potential for vegetation to affect slope stability through evapotranspiration depends primarily on vegetation type and the biogeoclimatic zone in question. Increases in air temperature generally intensify vapor pressure deficits, reduce the transfer of sensible heat from vegetation surfaces in humid areas, stimulate plant development, and lower the latent heat required to evaporate water. The overall effect is an increase in evapotranspiration (Allen *et al.*, 1991, Rosenzweig and Hillel, 1998). A sensitivity study conducted in a Kansas grassland revealed evapotranspiration increases ranging from 4 to 8% with a 1°C rise in mean air temperature (Rosenberg *et al.*, 1990).

Evapotranspiration has a seasonal influence on soil water budgets, which in turn influence the susceptibility of landsliding. During the rainy season, when landslides are most prevalent, evapotranspiration has little effect on antecedent soil moisture. Since most shallow, rapid landslides occur during periods of extended rainfall or snowmelt, evaporation is unlikely to be a controlling factor except during the first large rainstorm of the season (Megahan, 1983). However, in regions where winter snow accumulates, canopy structure plays a major role in intercepting snowfall (e.g., Golding and Swanson, 1986). Evaporative losses from snow intercepted by conifer forest canopies can be significant if snow persists in the canopy for many days; conversely, deciduous canopies trap little snow (Satterlund and Haupt, 1970). The influences of changing snow accumulation and vegetation types due to climate must be jointly evaluated to assess the importance of evapotranspiration on slope stability in areas with winter snowpacks.

Since deep-seated soil movements occur primarily during the wet season, evapotranspiration will not be a major factor in modifying movement rate in temperate regions (Sidle *et al.*, 1985). Increased evapotranspiration may slightly reduce the movement rate of slow, deep-seated mass movements in tropical and subtropical regions; however, this influence would be more important for decreasing the "window" of seasonal movement by modifying soil water conditions. Vegetation rooting depth in deep unstable regoliths is an important control on soil water depletion. Deeper-rooted vegetation species can sustain maximum transpiration rates for greater durations, thus drying the soil at greater depths compared to shallow-rooted vegetation (McNaughton and Jarvis, 1983). Therefore, climatic warming or human intervention that shifts vegetation from deep to shallow rooted species may increase the risk or movement rates of deep-seated mass failures. In parts of the world that experience a net increase in winter (or rainy season) precipitation related to climate change, the effect of the additional water on accelerating and initiating new deep-seated mass movements will undoubtedly overshadow any potential increases in evapotranspiration due to warming.

Rooting strength

Plant root systems enhance the stability of hillslopes by imparting a root cohesion component to the soil mantle, thus augmenting soil shear strength (Endo and Tsuruta, 1969; Wu *et al.*, 1979; Gray and Megahan, 1981; Abe and Ziemer, 1991). During storm or snowmelt periods when hillslope soils are in a tenuous state of equilibrium, reinforcement from tree roots may provide the critical difference between a stable and unstable site, especially when soils are partly or completely saturated (Sidle, 1992).

The stability of shallow hillslope soils is much more influenced by vegetation rooting strength than deeper soil mantles. In shallow soils, roots may penetrate the entire soil mantle and provide vertical anchors into the more stable substrate (Wu *et al.*, 1979; Gray and Megahan, 1981). Lateral roots distributed most densely in the upper soil horizons provide a membrane of strength that stabilizes the soil (Sidle *et al.*, 1985). This membrane strength is much more significant in ameliorating shallow rapid mass movements compared to deep-seated movements (Swanston and Swanson, 1976). Thus, vegetation root-strength changes associated with climatic warming would be expected to have a much greater effect on shallow, rapid landslides than on deep-seated mass movements. Surficial mass erosion (e.g., dry ravel) is highly influenced by surface roughness, partly due to the vegetation architecture (including the rigidity of the near-surface plant material, which depends on the surface rooting strength).

Climate-controlled studies of root biomass response for different agricultural species have concluded that in virtually all cases root dry weight increases under conditions of elevated atmospheric CO_2 (Rogers *et al.*, 1994). Such responses may also apply to forest species. Root:shoot ratios, a key indicator of morphological changes in response to CO_2, predominantly increased in controlled-climate experiments (Rogers *et al.*, 1994). Such responses might be attributed the reduction in the amount of nitrogen required to sustain maximum leaf photosynthetic rates in elevated CO_2 environments. Hence, additional nitrogen can be allocated to the roots for anabolic processes (Rosenzweig and Hillel, 1998). Since root biomass is correlated with root strength (Ziemer, 1981), this elevated CO_2-induced response may indicate greater cohesion due to denser and possibly stronger root systems. However, longer-term changes in vegetation community structure may be a more significant issue related to root strength and landslide susceptibility. As noted earlier, shifts from deciduous to coniferous species, from shrubs to grasslands, and from forest cover to shrubland will all tend to increase landslide potential based on changes in rooting strength. The reverse shifts of vegetation would, conversely, increase terrain stability. For dry ravel and dry creep, reductions in vegetation cover and shifts to a less rigid plant architecture promote increased rates (Sidle *et al.*, 1985, 1993).

Effects of Forest Management Practices

Different forest management practices such as timber harvesting, prescribed fire, roads, recreation, vegetation conversion, and agroforestry can affect hillslope stability in many ways (Sidle *et al.*, 1985). Although forest management activities may, in some cases, be flexible enough to partially mitigate the effects of climate change, assessment of the effects of forest management strategies on slope stability in a changing climate is very complicated. Additionally, from the viewpoint of broader global environmental change, we must consider anthropogenic impacts to forests that arise from changing management technologies, demographics, and social pressures and preferences. This portion of the chapter focuses on some of the key forest management issues that affect slope stability in the context of changing climates.

Timber harvesting

Timber harvesting can affect the stability of slopes in two ways: (i) temporarily increasing water inputs and soil moisture because of decreased evapotranspiration and changes in the volume and rate of snowmelt; and (ii) reducing root cohesion because of root strength deterioration (Sidle *et al.*, 1985). As noted earlier, the first factor is not particularly important for most landslides that occur during an extended rainy season. Decreased evapotranspiration after timber harvest could extend the "window of susceptibility" for landslide activity (Swanson and Swanston, 1977; Sidle *et al.*, 1985).

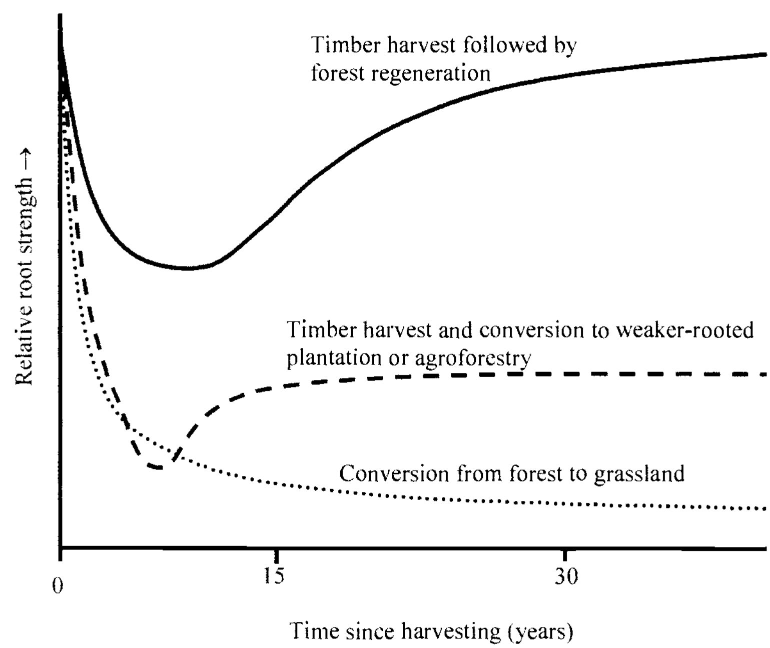


Fig. 8.5. Temporal changes in relative rooting strength in response to forest harvest and subsequent forest regeneration, conversion of forest to grassland, and conversion of natural forest to a weaker-rooted plantation or agroforestry. Minima in rooting strength represent windows of susceptibility for landsliding.

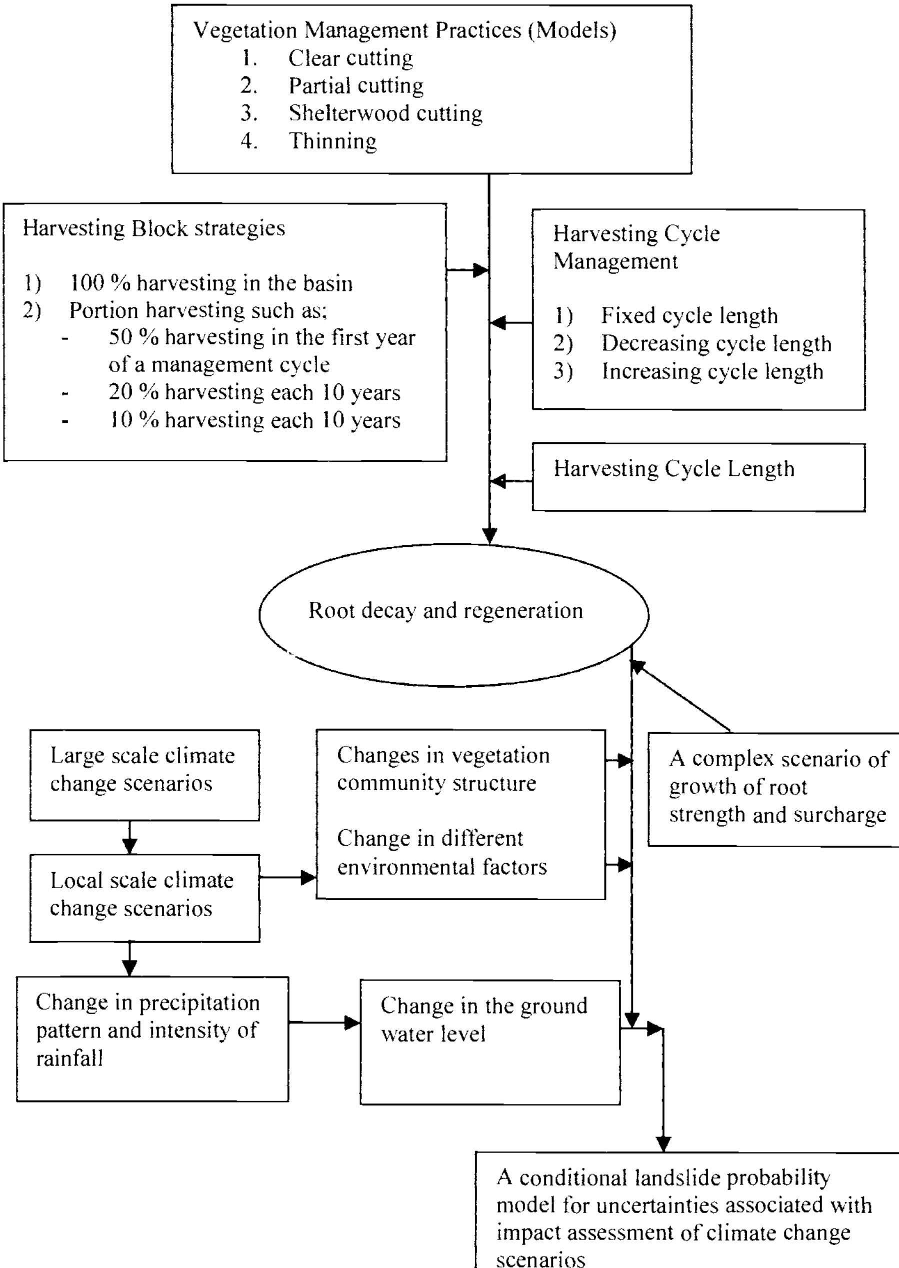


Fig. 8.6. Diagram of conditional probabilities for landslide response related to different climate change and timber harvest scenarios.

When large landslide-producing storms occur during drier conditions, decreased evapotranspiration in recently harvested sites could increase the susceptibility of landslide initiation. A more important factor linking timber harvesting with landslide initiation is the effect of deteriorating rooting strength after cutting. Increased frequency and volume of shallow landslides have been reported after forest harvesting by many investigators in mountainous terrain worldwide (Bishop and Stevens, 1964; Endo and Tsuruta, 1969; Swanston, 1970; Swanson and Dyrness, 1975; O'Loughlin and Pearce, 1976; Wu and Swanston, 1980; Ziemer, 1981; Wu and Sidle, 1995; Jakob, 2000). This increase typically occurs 3 to 15 years after forest harvesting, coinciding with the timing of minimum rooting strength at the site following initial root decay and prior to substantial regeneration (O'Loughlin and Pearce, 1976; Megahan *et al.*, 1978; Sidle, 1991; Wu and Sidle, 1995) (Fig. 8.5). Independent studies of the effects of timber harvest on rooting strength (Burroughs and Thomas, 1977; Ziemer and Swanston, 1977; Wu *et al.*, 1979; Abe and Ziemer, 1991) have confirmed these empirical observations. While these harvesting effects are more pronounced for shallow failure types, the loss of root strength may temporarily increase movement rates of soil creep and earthflows (Swanston and Swanson, 1976; Swanston, 1981).

To assess the effects of forest harvesting and forest conversion in the context of climate change, it is necessary to consider the influences of changes in vegetation community structure and the response of root systems to elevated CO_2 environments together with the impacts of harvesting and cover change. Thus, the predicted response of landslide activity in harvested and converted sites subject to climate change becomes a complex set of conditional probabilities (Fig. 8.6).

Prescribed fire/vegetation conversion

Prescribed fire is widely used in forest, rangeland and hillslope agricultural management as a mechanism for removing residues after cropping or timber harvesting. This practice can be used in the context of sustained management (e.g., burning forest brush prior to replanting trees) or conversion to another vegetation type (e.g., conversion of forest to pastureland). Controlled prescribed burning is also used in drier brushlands and forests to reduce wildfire hazard. Many soils exhibit water repellent properties after burning, especially in response to moderate to hot burns (DeBano *et al.*, 1967; Morris and Moses, 1987). These moderate to severe fires cause an immediate decrease in the infiltration capacity of surface soils which translates into increases in surface runoff and surface erosion and a possible decrease in landslide occurrence because of reduced water inputs into the soil mantle (Sidle *et al.*, 1985). However, as rooting systems decay and new vegetation restores infiltration capacities in 2 to 3 years, the burned areas become much more vulnerable to landsliding, similar to responses in clearcut forests (Rice, 1977). Rotational prescribed burning in both brushlands (Rice *et al.*, 1982) and heathlands

(Fairbairn, 1967) increased landslide rates several-fold during the period of minimum rooting strength (i.e., several years after the burn). Gray and Megahan (1981) observed a 20-fold increase in mass erosion after wildfire in granitic forest soils of central Idaho.

The role of fire in promoting landslides and debris flows is quite complicated and is frequently misrepresented. Former debris flow and landslide initiation sites in and around headwater channels fill in with sediment and organic material over a number of years or decades. The rates of such infilling processes are greatly enhanced after burning, particularly related to surficial transport of sediment and debris into these channels (Florsheim *et al.*, 1991; Wohl and Pearthree, 1991; Meyer *et al.*, 1992). Therefore, the rapid loading of these drainages that often occurs immediately after fire can increase the probability of a debris flow. Additionally, fires consume large wood in channels and thus may destabilize stored sediment and decrease channel roughness (Cannon and Reneau, 2000). However, a large rain event and subsequent high flow is needed to trigger a debris flow - these are the mass failures that typically occur shortly after fires (e.g., Wells, 1987; Cannon and Reneau, 2000).

Less perceptible but significant mass erosion immediately after fire occurs as dry ravel. Because much of the surface soil organic matter that binds soils together is destroyed during relatively hot fires, a significant portion of the interlocking frictional resistance among mineral grains or aggregates may be lost, thus lending these surface soils susceptible to gravitational movement immediately after burning (Sidle *et al.*, 1985; Florsheim *et al.*, 1991). Burning also removes the insulating organic cover from mineral soils making them more susceptible to freeze-thaw and wetting-drying cycles that in turn exacerbate dry ravel. Typically the effects of burning on dry ravel production are short-lived. Several studies have shown that the majority of dry ravel on steep, burned forest sites occurs almost immediately after the fire (Mersereau and Dyrness, 1972; Bennett, 1982). Rice (1982) measured dry ravel erosion of 39 m^3 ha^{-1} during a 3 month period after a fire in chaparral compared to more typical rates of about 1.4 m^3 ha^{-1} yr^{-1} in overall chaparral zones. As burned sites revegetate, dry ravel decreases markedly.

Fire has the potential to increase dry ravel and dry creep in the short term, increase debris flows in the short to moderate term, and increase landslides in the moderate to long term. Because the incidence and intensity of prescribed fire is largely controllable, it is the potential interaction between climate change and burning that may influence slope stability. Predicted increases in both storm frequency and intensity, in addition to average annual rainfall, indicate a higher probability for landsliding in the burned areas. Because climatic warming may cause decreases in protective organic soil matter in some regions, these sites could be predisposed to a greater risk of dry ravel just after burning. Additionally, higher diurnal and seasonal temperature fluctuations would promote more frequent wetting and drying and freeze-thaw cycles, thus increasing dry ravel on burned or disturbed sites. Longer, drier summers predicted in temperate climate change scenarios (e.g., Loaiciga *et al.*, 1996) would exacerbate wildfire hazard as well as the uncontrolled spread of a prescribed fire. These conditions would expand the spatial vulnerability to future fire-related mass erosion as well as increase the rates of ravel on burned sites.

Progressive forest clearing and conversion to other land covers are widespread environmental changes, especially during the past 50 years, in developing countries of Africa, Asia, and Latin America. The resulting deterioration of the land base is partly attributed to increases in mass wasting and surface erosion (Harwood, 1996; Fischer and Vasseur, 2000). Increases in landslide erosion following progressive forest conversion in Tanzania from the mid- to late 19th century are partly attributed to loss of rooting strength after clearing and burning (Haldemann, 1956). Widespread conversion of mixed evergreen forests to pasture following European settlement (1860-1920) in North Island, New Zealand caused severe erosion and reduced productivity of the land (Ministry of Works and Development, 1970; Trustrum *et al.*, 1983; Sidle *et al.*, 1985). Moderate to extreme mass movement erosion affected a total of 9280 km^2 of converted lands on North Island (Sidle *et al.*, 1985). In addition to these largely anthropogenic causes of forest conversion, climate change may also induce similar, but less drastic, changes in vegetation with similar consequences. A likely scenario for root-strength changes after conversion from forest to weaker-rooted vegetation is shown in Fig. 8.5.

Roads and trails

In steep forested terrain, road right-of-ways typically contribute at least an order of magnitude more landslide erosion per unit area disturbed compared to clearcut forests and 25 to 400 times more mass erosion than in undisturbed forests (Sidle *et al.*, 1985). Roads or trails established on any hillslope decrease the site stability in four ways: (i) overloading the slope in the embankment fill; (ii) oversteepening both the cut and fill slopes; (iii) removing support of the cutslope; and (iv) rerouting and concentrating road drainage water (Sidle *et al.*, 1985). Roads typically have a greater impact on shallow, rapid failures (Megahan and Kidd, 1972; Burroughs *et al.*, 1976), however, deep-seated mass failures can also be reactivated by cutting roads through the toe slope or placing material on the head of old slump blocks (Swanston and Swanson, 1976; Sidle *et al.*, 1985; Jones and Lee, 1989). Exposed substrate on steep cut and fill slopes is a source of dry ravel during wetting and drying cycles and freeze-thaw periods (Megahan, 1978; Sidle *et al.*, 1993).

Climate change may necessitate the construction of new roads to access forest stands that have shifted to higher elevations as the result of warming. Such higher elevation sites are often steeper, exacerbating landslide problems related to the expanded road system. Additionally, new tree species that emerge as the result of climate change may require different management systems and thus alterations in existing access roads. While such predictions may be somewhat speculative, the extent of the road-related slope stability problem (e.g., Sidle *et* *al.*, 1985) warrants further examination of climate change impacts. Unimproved trails in forests would be subject to similar impacts.

Recreation

Both the spatial extent and types of recreational activities on forestlands has changed dramatically in the last several decades. In developing countries, the increased popularity of mountainous backcountry trekking has greatly expanded former trail systems as well as introduced numerous new trails in unstable terrain. Additionally, recreational pressure has subjected trails formerly used only by people and animals to be widened for vehicles. In more developed nations, pressure from expansion of ski resorts in steep terrain (Fig. 8.7) and the expanding use of all-terrain vehicles has exacerbated slope instability (Watson, 1985). Within the past few decades there has been a rapid increase in the number of recreational and permanent dwellings in steep forest terrain. Such development along the Wasatch Front in Utah has exacerbated landslide and debris flow hazards. While these various recreational impacts have not been assessed to the extent of other land uses such as timber harvesting and roads, they are nonetheless quite significant.

Fig. 8.7. Mass erosion associated with a backcountry road into a nordic ski area. South Island, New Zealand.

Impacts of climate change related to landscape stability of recreation areas would involve many of the same considerations as with roads, trails, vegetation conversion, and timber harvesting. With climatic warming, forest clearing related to new ski slopes and other recreational uses may have to shift to higher elevation sites; such sites could be more susceptible to landslides. Considered in the broader context of global environmental change, new technology has brought to bear a series of unforeseen impacts on unstable hillslopes. Most of these impacts are in the privy of affluent residents and tourists or even ecotourists, but are nevertheless creating environmental damage as more and more people have access to steep and remote terrain. While certain land management agencies are paying lip service to these recreational impacts, the current philosophy of "customer satisfaction" and responsiveness to user needs practiced by the US Forest Service is avoiding this important environmental issue. Therefore, to expect better guidelines for protection against recreational impacts in developing countries is wishful thinking at best.

Agroforestry

Both indigenous and modern agroforestry systems rely heavily on soil management (Young, 1997). Shifting cultivation, the earliest form of agroforestry, had as one objective the restoration of soil fertility lost during intensive crop cultivation. However, shifting cultivation can accelerate soil mass movement due to the loss of rooting strength after clearing and burning forests (Lanly, 1969; Sidle *et al.*, 1985). Because shifting cultivation clears forested hillsides for long periods and because natural reforestation on abandoned, nutrient-depleted agricultural plots is very slow, the impact of this agroforestry practice on the stability of slopes may be much longer than the impact of timber harvesting (i.e., with regeneration) (Fig. 8.5).

Modern agroforestry practices involve land-use systems where trees are grown in association with agricultural crops, pasture, or livestock. Such agroforestry has the potential to contribute to environmental conservation by controlling land degradation and reducing the socio-economic pressures for forest clearance. The contemporary view is that modern agroforestry systems offer both socio-economic and biophysical benefits related to sustainability, however, there is little information on the actual impacts of such projects, especially at the larger scale (Steiner, 1988; Fischer and Vasseur, 2000).

It has been suggested that the wider adoption of agroforestry practices could alleviate the build-up of atmospheric CO_2 and other greenhouse gases (Dixon, 1995, 1996). Temporary increases in carbon storage may be achieved when more trees are grown on farmland. Widespread adoption of this practice would substantially increase soil organic matter as well as soil carbon storage. By reducing the pressure of widespread forest clearance through such agroforestry practices, reductions in atmospheric CO_2 can be achieved.

The magnitude of carbon sequestration attributed to agroforestry plantings will depend on the scale of implementation and the ultimate use of the wood. Large-scale agroforestry tree planting schemes such as the "Four Around" schemes in China have reportedly been implemented in more than 6.5 million ha of agricultural lands from 1980 to 1990 (Ciesla, 1995). A project of this magnitude provides numerous environmental benefits including the mitigation of shallow landsliding on hillsides, assuming that suitable forest cover establishes and is maintained.

Methods, Techniques and Applications of Landslide Hazard Assessment

Techniques and methodologies

Assessment techniques for landslide hazards can be roughly divided into three categories: (i) terrain stability classification (e.g., Kienholz, 1978; Ives and Messerli, 1981; Howes and Kenk, 1988; Rupke *et al.*, 1988; McKean *et al.*, 1991; Hearn, 1992); (ii) empirical landslide hazard assessment (e.g., Keefer *et al.*, 1987; Gupta and Joshi, 1990; Dhakal *et al.*, 2000); and (iii) physically-based models (e.g., Montgomery and Dietrich, 1994; Wu and Sidle, 1995) (Table 8.2). Each of these assessment techniques has some potential application for evaluating the impacts of climate change on slope stability.

In terrain stability classification, also called qualitative analysis or geomorphological analysis, numerous characteristics of terrain are used to define landforms. These data are typically assembled from aerial photographs, topographic maps, geological maps, and vegetation and ecosystem maps. Depending on the degree of detail required in the stability assessment, some field reconnaissance may be required (Table 8.2). The severity of the landslide hazard is then evaluated for portions of the terrain based on subjective decision rules. Such landslide mapping can be accomplished at levels ranging from reconnaissance (e.g., ≈ 1:20,000 to 1:50,000) to detailed (e.g., ≈ 1:5000 to 1:15,000) (Ministry of Forests, 1995).

In empirical landslide analyses, the causal and contributing factors are usually established based on the characteristics of existing landslides or on real-time landslide initiation data (e.g., rainfall). Empirical analyses are based on statistical, relational, and monitored methods (Table 8.2). In statistical analysis, the factors associated with topography, geology, and vegetation that are the primary indices of landslide susceptibility are quantified to assess their relative contributions to landslide initiation (Yin and Yan, 1988; Gupta and Joshi, 1990; Carrara *et al.*, 1991; Pachauri and Pant, 1992; Wang and Unwin, 1992; Van Westen, 1994; Chung *et al.*, 1995; Mark and Ellen, 1995; Sarkar *et al.*, 1995; Dhakal *et al.*, 1997, 1999). This approach assumes that landslides are more likely to occur under conditions similar to those of previous failures (Brabb, 1984; Varnes, 1984). Either a multivariate or univariate statistical approach may be applied, although a multivariate approach such as discriminant analysis (Davis, 1986) is considered to be better because it takes into account the interrelationships among factors. The relational methodology estimates landslide hazard from earlier and active landslide data by examining relationships between parameters such as slope angle and slope height (e.g., Zika *et al.*, 1988).

Table 8.2. Different methods of landslide hazard assessment.

Hazard assessment methods	Category	Sub-category	Methods of analysis	Hazard type	Type of data required	Appropriate analysis scale
Terrain stability classification method	Distribution analysis		Information on landslide distribution	Relative hazard	Aerial photographs and basic maps	All scales
	Geomorphological		Descriptive statistics from fields, qualitative map combination, etc.	Relative hazard	Maps, aerial photographs & extensive fieldwork	All scales
	Statistical	Univariate statistical analysis	Susceptibility analysis Information value method Failure rate analysis	Relative hazard	Maps, aerial photographs, considerable fieldwork	Medium
		Multivariate statistical analysis	Discriminant analysis Multiple regression analysis	Relative hazard	Maps, aerial photographs, considerable fieldwork	Medium
Empirical methods	Relational		Empirical curves	Empirical hazard	Past landslides and their relation to topographic features	Medium/ Large
	Monitored		Deformation monitoring	Monitored hazard	Detail information on triggering factors and landslidng	Medium
			Precipitation monitoring	Monitored hazard	Information on precipitation that caused landsliding	Medium
Physically based methods	Site specific analysis		Safety factor analysis	Absolute hazard	Detailed soil properties, pore pressure, slip surface etc.	Large (site specific)
	Basin based models		Distributed shallow landslide analysis model (dSLAM)	Absolute hazard	High resolution topographic data, soil, vegetation and rainfall data	Large/ Medium

Monitoring methods used in landslide hazard analysis include comparisons of real-time earthquake and rainfall records with landslide timing and distribution to establish threshold values for occurrence and severity of the hazard (e.g., Keefer, 1984; Cannon and Ellen, 1985; Keefer *et al.*, 1987; Capecchi and Focardi, 1988).

Physically based models (e.g., Skempton and Delory, 1957; Okimura, 1982; Sidle 1992; Montgomery and Dietrich, 1994; Terlien *et al.*, 1995) assess terrain stability in terms of a factor of safety (FS = the ratio of shear strength of the soil to shear stress) – for FS values >> 1 the slope is stable; as FS approaches 1 it becomes unstable; and for FS $\leq$ 1 slope failure occurs. The method is most suitable for smaller areas because in larger areas the variations in terrain and soil parameters included in the analysis of the safety factor are too large to be accurately quantified (Jibson and Keefer, 1989; Mulder, 1991) (Table 8.2). However, recent applications of such models have been able to capture the temporal and spatial effects of timber harvesting (as well as terrain parameters) on shallow landslide timing and volume (Wu and Sidle, 1995; Sidle and Wu, 1999).

Empirical landslide hazard assessment at the medium scale

An example of empirical landslide hazard assessment at the medium scale (1:25,000–1:50,000) that incorporates geographical information systems (GIS) for data analysis and display along with multivariate statistical analysis is briefly illustrated. The example is taken from a typical Himalayan mountain watershed, the Kulekhani watershed (124 km^2), located in the central region of Nepal.

A map showing the distribution of landslides was produced from aerial photo interpretation; afterwards, extensive field checking was completed. To determine the factors and classes (sub-factors) influencing landsliding, layers of topographic factors derived from a digital elevation model (DEM), geology, and land use/cover were analysed by discriminant (Quantification Scaling Type II; QS-II) analysis and the results used for hazard mapping. The sample of landslides that was surveyed in the field primarily guided the selection and classification of these factors. In many cases, factors considered are dependent upon availability of data. In this case, eight terrain factors were included: slope gradient, slope aspect, elevation, drainage basin order, distance from the ridge, distance from the valley, land use/cover, and geology. Each of these factors was divided into different classes ranging from three to seven based on field surveys that specify their relative contribution to landslide initiation.

The analysis requires sampling of landslide and non-landslide groups. The size of the smallest landslide or average size of the landslide can be used to determine the size of the analysis unit (i.e., grid-cells). The landslide grid-cells in this case represent about 45% of the total number of landslides. For each grid-cell of the landslide and non-landslide groups, class codes of the eight terrain factors are assigned for the Q-S II analysis. The remaining landslides (referred to as "test landslides") are later used for the evaluation of the hazard maps produced.

The Q-S II analysis is a multidimensional quantification analysis (Hayashi, 1950, 1954) that incorporates nominal data and is the same as discriminant analysis. The quantification is attained by using frequencies as input data to maximize the efficiency of discrimination (Hayashi, 1952, 1954). This method is suitable for landslide hazard assessment, because nominal variables (factors) such as geology or land use/cover are often most important to discriminate between landslide and non-landslide groups. Other discriminant functions (e.g., canonical) require interval or ratio data (Klecka, 1980). The linear Q-S II function (αa_q) for a sample belonging to a group q with n factors and m classes in a factor can be written as:

$$\alpha a_q = \sum_{j=1}^{n} \sum_{i=1}^{m} \delta a(ji)\, Xji$$

where, $\delta a(ji) = 1$, if sample a belongs to the I-th class of factor j, otherwise 0; Xji = score of the I-th class of factor j.

The quantification of classes of the factors (Xji) is done in such a way that the proportion of variance between the groups to the total variance (i.e., the correlation ratio, η^2), which takes the value between zero and one, is maximized. The degree of difference between the group means is measured by η. The efficiency of the discrimination is therefore given by η or η^2 (Hayashi, 1952, 1954). Since a large class score in a factor contributes more than a small one in the Q-S II functions, a class score and the range of scores of a factor (difference between the maximum and minimum scores of the classes) can be interpreted to determine their importance. For the data that are not sampled, the factor-classes are measured and the group to which they belong is predicted from the score of the classes. The scores can then be used to produce a map showing different relative classes of hazard (Dhakal *et al.*, 2000) (Fig. 8.8).

The incorporation of climatic change scenarios into such an empirical landslide hazard model is a complex task because many scenarios of climatic change are not yet clear. This hazard assessment method as well as most of the terrain stability classification systems and empirical landslide hazard models rely heavily on environmental factors (passive factors) and may or may not use causative factors (e.g., rainfall event data) to evaluate landslide hazard. To introduce the effect of climatic change in such methods, it is necessary to predict the response of environmental factors to climatic change. These factors may include: (i) prediction of land use changes; (ii) prediction of weathering regime; and (iii) prediction of changes in soil properties due to climatic change.

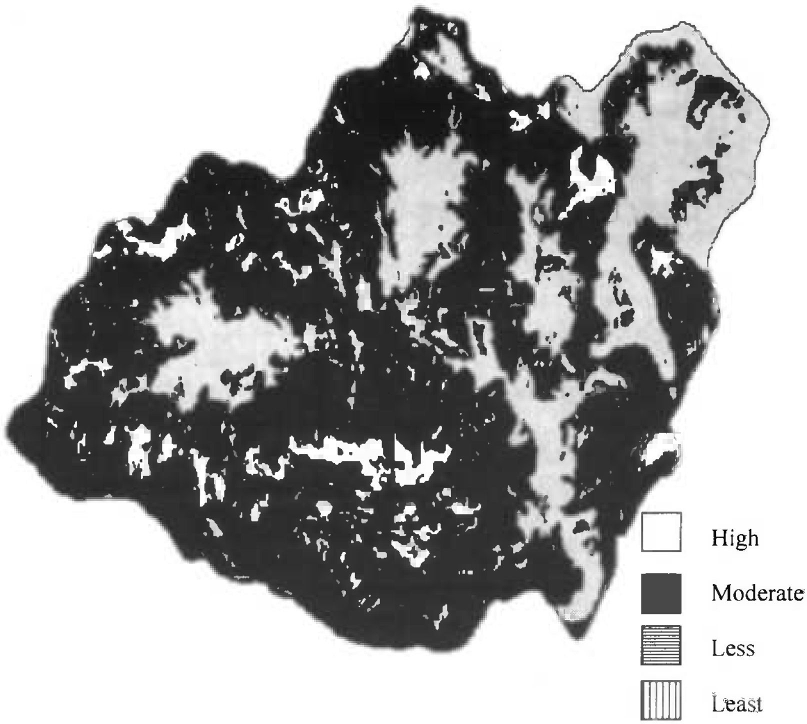


Fig. 8.8. Landslide hazard map of Kulekhani watershed in central Nepal; an example of medium-scale landslide hazard assessment and mapping. The multivariate statistical method (Q-SII analysis) employed for hazard zonation used eight different environmental factors to discriminate between the stable and unstable slopes.

Additionally, it is important to include 'surrogates' of specific triggering factors into landslide hazard analysis in order to capture the effects of future climate change scenarios. In a very general assessment of landslide hazard in the Lower Himalayas, Gupta and Joshi (1990) incorporated the distance of existing landslides from major tectonic features as a surrogate of earthquake triggering mechanisms. Precipitation indices or thresholds are surrogates that would be more closely linked to climate change. In a retrospective study that assessed the effects of climate change on landslide reactivation, Buma (2000) found that a semi-empirical model of net precipitation successfully predicted episodes of landslide movement based on a threshold of 3-month net antecedent precipitation. Net precipitation was estimated as liquid precipitation inputs minus evapotranspiration and soil water storage losses. Such parameters could be included on a seasonal basis in landslide models and modified according to plausible climate change scenarios to assess the potential impacts of climate change on landslide reactivation. Likewise, triggering mechanisms for shallow landslides such as rainfall and snowmelt, can be incorporated into landslide hazard analyses based on regional estimates of probabilities of rainfall intensity and total precipitation amounts and regional snowpack data. Expected changes in these precipitation inputs due to climate change could then be incorporated into such analyses.

However, our limited knowledge of future climate changes or trends based on GCM simulations does not provide an accurate basis for predicting the influence of either environmental factors (e.g., changes in soil weathering) or changes in the probability of triggering mechanisms. To significantly improve empirical landslide hazard assessments related to climate change, several significant issues need to be addressed: (i) methods that can be applied in broader geographic areas or in areas that experience multiple failure types (e.g., slump–earthflows, debris avalanches) need to be developed; (ii) there is a need to clearly focus on the underlying processes relating to slope failure; (iii) simple triggering thresholds for landslides must be identified that can be used in conjunction with crude climate change predictions; and (iv) the sensitivity of soils and vegetation to future climatic conditions must be addressed.

The first problem can partly be addressed by developing separate mapping rules for distinctly different failure types. For example, deep-seated slump-earthflow hazards would require different geologic, soils and hydrologic indicators compared to shallow rapid failures. The second issue is difficult to implement in data-sparse regions but as remotely sensed data become more available and accurate it may be possible to incorporate such important factors as rainfall distribution, detailed vegetation cover, digital elevation models (DEMs) and geomorphic features into hazard assessments. The third concern requires further regional-scale research on landslide triggering thresholds (e.g., rainfall, snowmelt) that are formulated in a meaningful way to relate to temporal restrictions of climate change forecasts. Finally, the fourth issue requires additional data to better quantify the likely response of soil and vegetation properties (that influence slope stability) to climate change. Additionally, the influences of forest operations and other land uses on slope stability need further investigation.

Application of a physically-based landslide model (dSLAM) at a detailed scale

A distributed, physically based model (dSLAM) developed to examine the spatial and temporal impacts of vegetation changes on shallow landslide probability is discussed in this example (Wu and Sidle, 1995; Sidle and Wu, 1999). The model incorporates an infinite slope analysis (based on factor of safety), continuous temporal changes in root cohesion and vegetation surcharge (Sidle, 1991, 1992), and the stochastic influence of rainfall on pore water pressure. The model accepts only rainfall inputs, which can be actual rainstorm records, rainfall events generated from historical data, or records synthesized using Monte Carlo techniques (Wu and Sidle, 1995; Sidle and Wu, 1999).

To apply the concept of distributed modelling to landslide analysis, the drainage basin must be partitioned into relatively homogeneous elements and then topographic parameters for each element are calculated or assigned. A stream tube model is adapted in this topographic analysis since it is consistent with the subsurface hydrologic and geomorphic processes (Wu and Sidle, 1995). In applying dSLAM to a small forest watershed (1.18 km^2) in coastal Oregon, it was assumed that Hortonian overland flow did not occur. Rainfall infiltrates rapidly into the forest soil and drains to streams as subsurface flow. The basin contains shallow to moderately shallow soils ($\approx$ 0.5-1.5 m) overlying sandstone bedrock. This low permeability bedrock acts as a potential failure plane for shallow rapid landslides and promotes the development of a transient shallow groundwater table. These groundwater levels are calculated in the model and are incorporated into the factor of safety equation to determine the dynamic stability of individual elements.

A series of rainstorms in the mid-1970s caused widespread landsliding in the area (especially in the clearcuts of the Douglas fir forest). A large, 21-hour storm (return interval 17-25 years) on November 29-30, 1975, which triggered many landslides in the region, was selected as the rainfall event for various landslide simulations (Wu and Sidle, 1995). Simulated volume (733 m^3) and numbers (4) of landslides in the watershed agreed closely with values (734 m^3 and 3, respectively) measured in the field after the 1975 storm. All simulated and actual landslides occurred in sites that were clearcut in 1968, 7 years prior to the November storm, thus linking landslide occurrence to the period of minimum rooting strength after logging at the site (Wu and Sidle, 1995). Locations of the simulated failures were in steep areas of groundwater convergence, particularly geomorphic hollows or zero-order basins (Fig. 8.9).

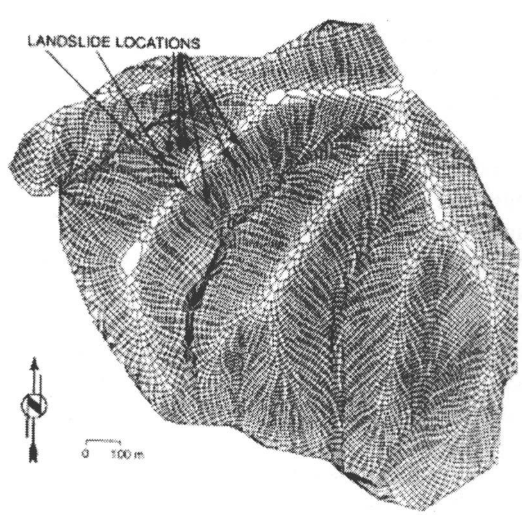


Fig. 8.9. Landslide locations during a November 1975 storm simulated by dSLAM for the Cedar Creek Watershed, Oregon Coast Ranges, USA.

Monte Carlo simulations were also conducted to assess the effects of various timber harvesting scenarios in the watershed (Sidle and Wu, 1999). For a single 50% clearcut, most simulated landslides were distributed in a period of about 3 to 15 years after the clearcut. The spatial distribution of failure probability in the Cedar Creek basin was investigated for several cases: the actual pattern of timber harvesting in 1975; a random 50% clearcut with and without leave areas at the beginning of the cycle; and a 100% clearcut at the beginning of the cycle with and without leave areas. Vegetation leave areas, a forest management practice employed to reduce landslide occurrence in unstable portions of proposed clearcuts, was evaluated to be an effective control measure.

Physically based landslide models such as dSLAM appear to be powerful tools to assess future scenarios of climate change affecting slope stability. The following factors related to climate change can easily be incorporated into dSLAM: (i) estimates of changes in precipitation intensity and volume (e.g., a 'design' storm); (ii) estimates of changes in antecedent soil moisture conditions; (iii) estimates of changes in root cohesion due to introduction of different species; and (iv) long-term estimates of changes in soil properties (e.g., depth, cohesion) due to increased weathering. The most efficient way to evaluate these complex interactions would be to develop long-term Monte Carlo simulations for selected small catchments in regions where climate responses are expected. The modelling would be a formidable task and would be limited primarily by the estimates of climate change parameter responses, especially rainfall intensity and total storm precipitation. Typical climate change scenarios of ± 20% parameter response could initially be used to estimate the sensitivity of landslides to hypothetical changes.

Discussions and Conclusions

Climate is a significant forcing mechanism for many geomorphological processes. The consequences of climatic change related to landslide magnitudes and frequencies are important issues for future landslide predictions. However, the incorporation of climatic change scenarios into landslide hazard models is a complex task because scenarios of climatic change are not yet clear even at regional levels. The prediction complexity further increases when we try to relate the influence of projected temperature rises to environmental factors such as species composition changes, rooting depth, rooting strength, evapotranspiration, soil depth, soil cohesion, and vegetation regrowth characteristics. Temperature change in itself will affect weathering rates (Chigira, Chapter 7, this volume). Similarly, the distribution and vertical zonation of the major plant communities will adjust to altered climate conditions. Changes in vegetation in turn will affect weathering rates, soil characteristics, and freeze-thaw cycles among other things. Future precipitation patterns are the most critical factors affecting landslides. Probable scenarios of rainfall intensity and duration and snowmelt timing and magnitude that influence landslide initiation or reactivation are weak at best (Loaiciga *et al.*, 1996; Buma and Dehn, 1998).

Depending on the level of detail required and type of available data, landslide hazard can be analysed by several methods. Such hazard assessments are, in themselves, complicated tasks and require subjective decisions at various stages. Consequently, results are influenced to some extent by the methods employed and the skill of the user. GIS technology has greatly enhanced the capabilities of landslide modelling at all scales (Gupta and Joshi, 1990; Carrara *et al.*, 1991; Wu and Sidle, 1995; Dhakal *et al.*, 1999). GIS can also facilitate a quick trial and error approach for assessment methods and is an important tool for incorporating various analytical aspects.

The incorporation of climatic change scenarios into terrain stability classification and empirical landslide hazard assessments seem to be a complex task given the present level of understanding about climatic change scenarios. These methods often rely heavily on environmental factors (passive factors), and our limited knowledge of future climate changes or trends based on GCM simulations does not provide accurate prediction of changes in these environmental factors.

For physically based models (e.g., dSLAM), it is possible to incorporate projected future precipitation directly into the model. At present, different climatic models such as GCMs are available which simulate air pressure to produce future precipitation. It is possible that such derived precipitation can be used in the hydrological models to calculate the pore pressures and then the factor of safety. However, the major shortcoming of such climatic projection methods is their low horizontal resolution, currently around 250 × 250 km. The distributed hydrological models for slope stability generally need to be implemented at far more detailed scales – in the order of square kilometres to several tens of square kilometres. Downscaling techniques have been applied with some success to precipitation data to transfer GCM results to the local scale (e.g., Marinucci and Giorgi, 1992; Frey-Buness *et al.*, 1995; Heyen *et al.*, 1996). Problems with such approaches related to landslide assessment are inaccuracies of temporal distribution of precipitation and failure to capture the spatial and temporal responses associated with geomorphic process (e.g., Buma and Dehn, 1998; Dehn and Buma, 1999). Thus, it appears optimistic to consider current downscaling techniques for generating precipitation to be useful in physically based landslide models.

If climatic change models improve to the point where they can accurately predict changes in rainfall at the local scale, the incorporation of such projected future rainfall scenarios into the physically based models will be possible. However, similar predictions of changes in different environmental factors (e.g., soil and vegetation) are problematic. Best estimates of complex climate change responses of these factors may have to be incorporated into physically based models.

References

Abe, K. and Ziemer, R.R. (1991) Effect of tree roots on a shear zone: modeling reinforced shear stress. *Canadian Journal of Forest Research* 21, 1012-1019.

Acosta, R., Allen, M., Cherian, A., Granich, S., Mintzer, I., Suarez, A. and von Hippel, D. (1999) *Climate Change Information Sheets.* United Nations Environment Programme Information Unit for Conventions (UNEP/IUC), Châtelaine, Switzerland, 62 pp.

Allen, R.G., Gichuiki, F.N. and Rosenzweig, C. (1991) CO_2-induced climatic changes and irrigation-water requirements. *ASCE Journal of Water Resources Planning and Management* 117, 157-178.

Anderson, M.G. and Burt, T.P. (1977) Automatic monitoring of soil moisture conditions in a hillslope spur and hollow. *Journal of Hydrology* 33, 27-36.

Aniya, M. (1985) Landslide-susceptibility mapping in the Amahata river basin, Japan. *Annals of the Association of American Geographers* 75, 102-114.

Asare, S.N., Rudra, R.P., Dickinson, W.T. and Wall, W.T. (1997) Frequency of freeze-thaw cycles, bulk density and saturation effects on soil surface shear and aggregate stability in resisting water erosion. *Canadian Agricultural Engineering* 39, 273-279.

Bangxing, T., Suqing, L. and Shijian, L. (1994) Mountain disaster formation in northwest Sichuan. *GeoJournal* 34, 41-46.

Bechini, C. (1993) Natural conditions controlling earthflows occurrence in the Eden Canyon Area (San Francisco Bay Region, California). *Zeistschrift für Geomorphologie* Suppl. Bd. 87, 91-105.

Bello, N.J. (1998) Evidence of climate change based on rainfall records in Nigeria. *Weather* 53, 412-418.

Beniston, M. and Innes, J.L. (1998) Impacts of climatic variability and extreme on forests: synthesis. In: Beniston, M. and Innes, J.L. (eds), *The Impacts of Climate Variability on Forests.* Lecture Notes in Earth Sciences No. 74, Springer, Berlin, pp. 309-318.

Bennett, K.A. (1982) Effects of slash burning on surface soil erosion rates in the Oregon Coast Range. Unpublished MS thesis, Oregon State University, Corvallis.

Bishop, D.M. and Stevens, M.E. (1964) Landslides on logged areas, southeast Alaska. *USDA Forest Serv. Res. Rep. NOR-1,* Juneau, Alaska, 18 pp.

Boer, M.M., Koster E.A. and Lundberg, H. (1990) Greenhouse impacts in Fennoscandia – preliminary findings of a European workshop on the effects of climate change. *Ambio* 19, 2-10.

Bovis, M.J. and Jones, P. (1992) Holocene history of earthflow mass movements in south-central British Columbia: the influence of hydroclimatic changes. *Canadian Journal of Earth Sciences* 29, 1746-1755.

Brabb, E.E. (1984) Innovative approach to landslide hazard and risk mapping. In: *Proceedings of the 4th International Symposium on Landslides, Toronto, Ontario, Canada.* Vol. 1, pp. 307-324.

Buma, J. (2000) Finding the most suitable slope stability model for the assessment of the impact of climate change on a landslide in southeast France. *Earth Surface Processes and Landforms* 25, 565-582.

Buma, J. and Dehn, M. (1998) A method for predicting the impact of climate change on slope stability. *Environmental Geology* 35, 190-196.

Burroughs, E.R. and Thomas, B.R. (1977) Declining root strength in Douglas-fir after felling as a factor in slope stability. *USDA Forest Service Research Paper INT-190,* Intermountain Forest and Range Experiment Station, Ogden, UT.

Burroughs, E.R., Chalfant, G.R. and Townsend, M.A. (1976) *Slope Stability in Road Construction.* USDI, Bureau of Land Management, Portland, Oregon.

Campbell, A.P. (1966) Measurement of movement of an earthflow. *Soil Water* 2, 23-24.

Campbell, G.S. (1985) *Soil Physics with Basic Transport Models for Soil-Plant Systems.* Developments in Soil Science 14, Elsevier, Amsterdam.

Cannon, S.H. and Ellen, S.D. (1985) Rainfall conditions for abundant debris avalanches, San Francisco Bay region, California. *California Geology* 38, 267-272.

Cannon, S.H. and Reneau, S.L. (2000) Conditions for generation of fire-related debris flows, Capulin Canyon, New Mexico. *Earth Surface Processes and Landforms* 25, 1103-1121.

Capecchi, F. and Focardi, P. (1988) Rainfall and landslides: Research into a critical precipitation coefficient in an area of Italy. In: *Proceedings of the 4th International Symposium on Landslides, Lausanne.* Vol. 2, pp. 1031-1136.

Carrara, A., Cardinali, M., Detti, R., Guzzetti, F., Pasqui, N. and Reichenbach, P. (1991) GIS technique and statistical models in evaluating landslide hazard. *Earth Surface Processes and Landforms* 16, 427-445.

Cess, R.D., Potter, G.L., Blanchet, J.P., Boer, G.J., Del Genio, A.D., Déqué, M., Dymnikov, V., Galin, V., Gates, W.L., Ghan, S.J., Kiehl, J.T., Lacis, A.A., Le Treut, H., Li, Z.-X., Liang, X.-Z., McAvaney, B.J., Meleshko, V.P., Mitchell, J.F.B., Morcrette, J.-J., Randall, D.A., Rikus, L., Roeckner, E., Royer, J.F., Schlese, U., Sheinin, D.A., Slingo, A., Sokolov, A.P., Taylor, K.E., Washington, W.M., Wetherald, R.T., Yagai, I. and Zhang, M.-H. (1990) Intercomparison and interpretation of climate feedback processes in 19 atmospheric general circulation models. *Journal of Geophysical Research* 95(D10), 16, 601-16, 615.

Chung, C.F., Fabbri, A.G. and Van Westen, C.J. (1995) Multivariate regression analysis for landslide hazard zonation. In: Carrara, A. and Guzzetti, F. (eds), *Geographical Information Systems in Assessing Natural Hazards.* Kluwer Academic Publishers, Dordrecht, pp. 107-134.

Ciesla, W.M. (1995) Climate Change, Forests and Forest Management, An Overview. *FAO Forestry Paper 126*, FAO, United Nations, Rome.

Cubasch, U., Waszkewitz, J., Hegerl, G.C. and Perlwitz, J. (1995) Regional climate changes as simulated in time-slice experiments. *Climate Change* 31, 273-304.

Davis, J.C. (1986) *Statistics and Data Analysis in Geology.* John Wiley & Sons, New York.

DeBano, L.F., Osborn, J.F., Krammes, J.S. and Letey, J. (1967) Soil wettability and wetting agents... our current knowledge of the problem. *USDA Forest Service Research Paper PSW-43*, Berkeley, California.

Dehn M. and Buma, J. (1999) Modelling future landslide activity based on general circulation models. *Geomorphology* 30, 175-187.

Dhakal, A.S., Amada, T. and Aniya, M. (1997) A GIS approach to landslide susceptibility mapping: A case study from the Kulekhani watershed, Nepal. In: *Proceedings from 18th Asian Conference on Remote Sensing, Kuala Lumpur, Malaysia.* pp. JS-6-1 to JS-6-6.

Dhakal, A.S., Amada, T. and Aniya, M. (1999) Landslide hazard mapping and the application of GIS in the Kulekhani watershed Nepal. *Mountain Research and Development* 19, 3-16.

Dhakal, A.S., Amada, T. and Aniya, M. (2000) Landslide hazard mapping and its evaluation using GIS: An investigation of sampling scheme for grid-cell based quantitative method. *Photogrammetric Engineering and Remote Sensing* 66, 981-989.

Dietrich, W.E. and Dunne, T. (1978) Sediment budget for a small catchment in mountainous terrain. *Zeitschrift für Geomorphologie Suppl. Bd.* 29, 191-206.

Dingman, S.L. (1994) *Physical Hydrology.* Macmillan, New York.

Dixon, R.K. (1995) Agroforestry systems and greenhouse gases. *Agroforestry Today* 8, 11-14.

Dixon, R.K. (1996) Agroforestry systems: sources or sinks of greenhouse gases? *Agroforestry Systems* 31, 99-116.

Easterling, D.R., Diaz, H.F., Douglas, A.V., Hogg, W.D., Kunkel, K.E., Rogers, J.C. and Wilkinson, J.F. (1999) Long-term observations for monitoring extremes in the Americas. *Climatic Change* 42, 285-308.

Eckholm, E. (1979) Planning for the future: Forestry for human needs. *Worldwatch Paper 26*, Worldwatch Inst., Washington, DC.

Eisbacher, G.H. and Clague, J.J. (1984) Destructive mass movements in high mountains: hazard and management. *Geological Survey of Canada Paper 84-16*, Ottawa, Canada.

Endo, T. and Tsuruta, T. (1969) The effect of the tree's roots on the shear strength of soil. *Annual Report, 1968, Hokkaido Branch For. Exp. Stn.* Sapporo, Japan, pp.167-182.

Evans, S.G. and Clague, J.J. (1994) Recent climate change and catastrophic geomorphic processes in mountain environments. *Geomorphology* 10, 107-128.

Fairbairn, W.A. (1967) Erosion in the River Findhorn Valley. *Scottish Geograph. Mag.* 83, 46-52.

Fischer, A. and Vasseur, L. (2000) The crisis in shifting cultivation practices and the promise of agroforestry: a review of the Panamanian experience. *Biodiversity and Conservation* 9, 739-756.

Florsheim, J.L., Keller, E.A. and Best, D.W. (1991) Fluvial sediment transport in response to moderate storm flows following chaparral wildfire, Ventura County, southern California. *Geological Society of America Bulletin* 103, 504-511.

Francis D. and Hengeveld, H. (1998) *Extreme Weather and Climate Change.* Environment Canada, Ministry of Supply and Services, Canada.

Frey-Buness, A., Heimann, D. and Sausen, R. (1995) A statistical-dynamical downscaling procedure for global climate simulations. *Theoretical Applied Climatology* 50, 117-131.

Fu, C. and Wen, G. (1999) Variation of ecosystems over East Asia in association with seasonal, interannual and decadal monsoon climate variability. *Climatic Change* 43, 477-494.

Golding, D.L. and Swanson, R.H. (1986) Snow distribution patterns in clearings and adjacent forest. *Water Resources Research* 22, 1931-1940.

Gray, D.H. and Megahan, W.F. (1981) Forest vegetation removal and slope stability in the Idaho Batholith. *USDA Forest Service Research Paper INT-271*, Ogden, Utah.

Gregory, K.J. and Walling, D.E. (1973) *Drainage Basin Form and Process.* John Wiley, New York.

Gupta, R.P. and Joshi, B.C. (1990) Landslide hazard zoning using the GIS approach – A case study from the Ramganga catchment, Himalayas. *Engineering Geology* 28, 119-131.

Haeberli, W. and Burn, C. (2002) Natural hazards in forests: glacier and permafrost effects as related to climate change. In: Sidle, R.C. (ed.), *Environmental Change and Geomorphic Hazards in Forests.* IUFRO Research Series, CAB International, Wallingford, UK, pp. 167-202.

Haldemann, E.G. (1956) Recent landslide phenomena in the Rungwe volcanic area, Tanganyika. *Tanganyika Notes Rec.* 45, 3-14.

Harwood, R.R. (1996) Development pathways toward sustainable systems following slash-and-burn. *Agriculture, Ecosystems and Environment* 58, 75-86.

Hayashi, C. (1950) On the quantification of qualitative data from the mathematico-statistical point of view. *Annals of the Institute of Statistical Mathematics* 1, 35-47.

Hayashi, C. (1952) On the prediction of phenomena from qualitative data and the quantification of qualitative data from the mathematico-statistical point of view. *Annals of the Institute of Statistical Mathematics* 3, 69-98.

Hayashi, C. (1954) Multidimensional quantification – with the applications to analysis of social phenomena. *Annals of the Institute of Statistical Mathematics* 5, 121-143.

Hearn, G.J. (1992) Terrain hazard mapping at Ok Tedi mine, Papua New Guinea. In: *Proceedings of 6th International Symposium on Landslides, Christchurch, New Zealand.* Vol. 2, pp. 971-976.

Heyen, H., Zorita, E. and von Storch, H. (1996) Statistical downscaling of monthly mean North Atlantic air-pressure to sea level anomalies in the Baltic Sea. *Tellus* 48A, 312-323.

Hough, B.K. (1951) An analysis of the effect of particle interlocking on the strength of cohesionless soils. *American Society Testing and Materials Bulletin* 176.

Howes, D.E. and Kenk, E. (1988) Terrain classification system for British Columbia (revised edn), *Ministry of Environment Manual 10*, Ministry of Crown Lands, Victoria, BC, Canada.

Hulme, M. and Viner, D. (1998) A climate change scenario for the tropics. *Climatic Change* 39, 145-176.

Intergovernmental Panel on Climate Change (IPCC) (1990) *Climate change, the IPCC scientific assessment.* Houghton, J.T., Jenkins, G.J. and Ephraums, J.J. (eds), Cambridge University, Cambridge, UK.

Ives, J.D. and Messerli, B. (1981) Mountain Hazards Mapping in Nepal. Introduction to an applied mountain research project. *Mountain Research and Development* 1, 223-230.

Jakob, M. (2000) The impacts of logging on landslide activity at Clayoquot Sound, British Columbia. *Catena* 38, 279-300.

Jibson, R.W. and Keefer, D.K. (1989) Statistical analysis of factors affecting landslide distribution in the New Madrid seismic zone Tennessee and Kentucky. *Engineering Geology* 27, 509-542.

Jones, J.A.A. (1997) *Global Hydrology: Processes, Resources and Environmental Management.* Longman, Essex, UK.

Jones, D.K.C. and Lee, E.M. (1989) The Catak landslide disaster, Turkey. *Quarterly Journal of Engineering Geology, London* 22, 93-95.

Keefer, D.K. (1984) Landslides caused by earthquakes. *Geological Society of America Bulletin* 95, 406-421.

Keefer, D.K., Wilson, R.C., Mark, R.K., Brabb, E.E., Brown, W.M., Ellen, S.D., Harp, E.L., Wieczorek, G.F., Alger, C.S. and Zatkin, R.S. (1987) Real-time landslide warning during heavy rainfall. *Science* 238, 921-925.

Kelsey, H.M., Madej, M.A., Pitlick, J., Stroud, P. and Coghlan, M. (1981) Major sediment sources and limits to the effectiveness of erosion control techniques in highly erosive watersheds of north coastal California. *IAHS Publication No. 132*, pp. 493-509.

Kienholz, H. (1978) Maps of geomorphology and natural hazards of Grindelwald, Switzerland, scale 1:10,000. *Arctic and Alpine Research* 10, 169-184.

Klecka, R.W. (1980) *Discriminant Analysis.* Sage University Paper Series Quantitative Applications in the Social Sciences, series no. 07-019, Beverly Hills and London: Sage Publications, 71 pp.

Krammes, J.S. (1965) Seasonal debris movement from steep mountainside slopes in southern California. In: *Proceedings of the Federal Inter-agency Sedimentation Conference. Jackson, Miss.* USDA Miscellanous Publication 970, pp. 85-88.

Lamberti, G.A., Gregory, S.V., Ashkenas, L.R., Wildman, R.C. and Moore, K.M.S. (1991) Stream ecosystem recovery following a catastrophic debris flow. *Canadian Journal of Fisheries and Aquatic Sciences* 48, 196-208.

Landsea, C.W., Pielke, R.A., Mestas-Nuñez, A.M. and Knaff, J.A. (1999) Atlantic basin hurricanes: indices of climatic changes. *Climatic Change* 42, 89-129.

Lanly, J.P. (1969) Regression de la foret dense en Cote d'Ivoire. *Bois et Forets des Tropiques* 127, 45-59.

Li, T. (1989) Landslides: extent and economic significance in China. In: Brabb, E.E. and Harrod, B.L. (eds), *Landslides: Extent and Economic Significance. Proceedings of the 28th International Geological Congress*, Washington, DC. A.A. Balkema, Rotterdam, The Netherlands, pp. 271-287.

Lindzen, R.S. (1990) Some coolness concerning global warming. *Bulletin of American Meteorological Society* 71, 288-299.

Loaiciga, H.A., Valdes, J.B., Vogel, R., Garvey, J. and Schwarz, H. (1996) Global warming and the hydrologic cycle. *Journal of Hydrology* 174, 83-127.

Manabe, S. and Weatherald, R.T. (1987) Large-scale changes in soil wetness induced by an increase in carbon-dioxide. *Journal of Atmospheric Science* 44, 1211-1235.

Mansell, M.G. (1997) The effect of climate change on rainfall trends and flooding risk in the west of Scotland. *Nordic Hydrology* 28, 37-50.

Marinucci, M.R. and Giorgi, F. (1992) A $2\times CO_2$ climate change scenario over Europe generated using a limited area model nested in a general circulation model. 1. Present-day seasonal climate simulation. *Journal of Geophysical Research* 97, 9989-10,009.

Mark, R.K. and Ellen, S.D. (1995) Statistical and simulation models for mapping debris-flow hazard. In: Carrara, A. and Guzzetti, F. (eds), *Geographical Information Systems in Assessing Natural Hazards*. Kluwer Academic, Dordrecht, pp. 93-106,

Mason, S.J., Waylen, P.R., Mimmack, G.M., Rajaratham, B. and Harrison, J.M. (1999) Changes in extreme rainfall in South Africa. *Climatic Change* 41, 249-257.

McKean, J., Buechel, S. and Gaydos, L. (1991) Remote sensing and landslide hazard assessment. *Photogrammetric Engineering and Remote Sensing* 57, 1185-1193.

McNaughton, K.G. and Jarvis, P.G. (1983) Predicting effects of vegetation changes on transpiration and evaporation. In: Kozlowski, T.T. (ed.), *Water Deficits and Plant Growth*. Vol. 7. Academic, New York, pp. 1-47.

Megahan, W.F. (1978) Erosion processes on steep granitic road fills in central Idaho. *Soil Science Society of America Journal* 43, 350-357.

Megahan, W.F. (1983) Hydrologic effects of clearcutting and wildlife on steep granitic slopes in Idaho. *Water Resources Research* 19, 811-819.

Megahan, W.F. and Kidd, W.J. (1972) Effect of logging and logging roads on erosion and sediment deposition from steep terrain. *Journal of Forestry* 70, 136-141.

Megahan, W.F., Day, N.F. and Bliss, T.M. (1978) Landslide occurrence in the western and central northern Rocky Mountain physiographic province in Idaho. In: *Proceedings of 5th North American Forest Soils Conference, Colorado State University, Fort Collins*. pp. 116-139.

Mersereau, R.C. and Dyrness, C.T. (1972) Accelerated mass wasting after logging and slash burning in western Oregon. *Journal of Soil and Water Conservation* 27, 112-114.

Meyer, G.A., Wells, S.G., Balling, R.C. and Jull, A.J.T. (1992) Response of alluvial systems to fire and climate change in Yellowstone National Park. *Nature* 357, 147-149.

Mimikou, M., Kouvopoulos, Y., Cavadias, G. and Vayianos, N. (1991) Regional hydrological effects of climate change. *Journal of Hydrology* 123, 119-146.

Ministry of Forests (1995) *Mapping and assessing terrain stability guidebook.* British Columbia Ministry of Forests, Victoria, BC, Canada.

Ministry of Works and Development (1970) Wise Land Use and Community Development. *Report of technical committee of inquiry into the problems of the Poverty Bay-East Cape District of New Zealand.* Wellington, NZ.

Monteith, J.L. (1976) *Vegetation and the Atmosphere,* Vol.2. Academic, New York.

Montgomery, D.R. and Dietrich, W.E. (1994) A physically based model for the topographic control on shallow landsliding. *Water Resources Research* 30, 1153-1171.

Morris, S.E. and Moses, T.A. (1987) Forest fire and the natural soil erosion regime in the Colorado Front Range. *Annals of the Association of American Geographer* 77, 245-254.

Mulder, H.F.H.M. (1991) *Assessment of Landslide Hazard.* Faculty of Geographical Sciences, University of Utrecht, The Netherlands.

Naranjo-Diaj, L.R. and Centella, A. (1998) Recent trends in the climate of Cuba. *Weather* 53, 78-81.

Okimura, T. (1982) Situation of surficial slope failure based on the distribution of soil Layer. *Shin-sabo* 35, 9-18 (in Japanese with English abstract).

O'Loughlin, C.L. and Pearce, A.J. (1976) Influence of Cenozoic geology on mass movement and sediment yield response to forest removal, North Westland, New Zealand. *Bulletin of International Association of Engineering Geology* 14, 41-46.

Onda, Y. and Matsukura, Y. (1997) Mechanism for the instability of slopes composed of granular materials. *Earth Surface Processes and Landforms* 22, 401-411.

Onda, Y., Mori, A. and Shindo, S. (1992) The effects of topographic convergence and location of past landslides on subsurface water movement on granite hillslope. *Journal of Natural Disaster Science* 14, 45-58.

Oyagi, N. (ed.) (1977) *Guide-book for Excursions of Landslides in Central Japan.* Japan Society of Landslides, Tokyo.

Pachauri, A.K. and Pant, M. (1992) Landslide hazard mapping based on geological attributes. *Engineering Geology* 32, 81-100.

Panagoulia, D. and Dimou, G. (1997) Linkage space-time scale in hydrological modelling with respect to global climate change, Part 2. Hydrological response for alternative climates. *Journal of Hydrology* 194, 39-63.

Pearce, A.J. and Watson, A.J. (1983) Medium-term effects of two landsliding episodes on channel storage of sediment. *Earth Surface Processes and Landforms* 8, 29-39.

Peterson, D.L. (1998) Climate, limiting factors and environmental change in high-altitude forests of Western North America. In: Beniston, M. and Innes, J.L. (eds), *The Impacts of Climate Variability on Forests.* Lecture Notes in Earth Sciences No. 74, Springer, Berlin, 191-208.

Pfister, L., Humbert, J. and Hoffman, L. (2000) Recent trends in rainfall-runoff characteristics in the Alzette River Basin, Luxembourg. *Climatic Change* 45, 323-337.

Rahn, P.H. (1969) The relationship between natural forested slopes and angles of repose for sand and gravel. *Geological Society of America Bulletin* 80, 2123-2128.

Reifsnyder, W.E. and Lull, H.W. (1965) Radiant Energy in Relation to Forests. *USDA Technical Bulletin No. 1344*, Forest Service, Washington, DC.

Reneau, S.L. and Dietrich, W.E. (1991) Erosion rates in the southern Oregon Coast Range: evidence for an equilibrium between hillslope erosion and sediment yield. *Earth Surface Processes and Landforms* 16, 307-322.

Rice, R.M. (1977) Forest management to minimize landslide risk. In: Kunkle, S.H. and Thames, J.L. (eds), *Guidelines for Watershed Management.* FAO, United Nations, Rome, Italy, pp. 271-289.

Rice, R.M. (1982) Sedimentation in the chaparral: How do you handle unusual events? *Sediment Budgets and Routing in Forested Drainage basins. USDA Forest Service General Technical Report PNW-141*, Portland, Oregon, pp. 39-49.

Rice, R.M., Ziemer, R.R. and Hankin, S.C. (1982) Slope stability effects of fuel management strategies: Inferences from Monte Carlo simulations. *USDA Forest Service General Technical Report PSW-58*, Berkeley, California, pp. 365-371.

Rogers, H.H., Runion, G.B. and Krupa, S.V. (1994) Plant responses to atmospheric CO_2 enrichment with emphasis on roots and the rhizosphere. *Environmental Pollution* 83, 155-189.

Rosenberg, N.J., Kimball, B.A., Martin, P. and Cooper, C.F. (1990) From climate and CO_2 enrichment to evapotranspiration. In: Waggoner, P.E. (ed.), *Climate Change and US Water Resources.* Wiley, New York, pp. 177-206.

Rosenzweig, C. and Hillel, D. (1998) *Climate Change and The Global Harvest- Potential Impacts of the Greenhouse Effect on Agriculture.* Oxford University Press, New York.

Rowntree, P.R. (1993) Climatic models – changes in physical environmental conditions, In: Atkinson, D. (ed.), *Global climate change its implications for crop protection.* BCPC Monograph No. 56, British Crop Protection Council, Surrey, UK, pp. 13-32.

Rupke, J., Cammeraat, E., Seijmonsbergen, A.C. and Van Westen, C.J. (1988) Engineering geomorphology of the Widentobel catchment, Appenzell and Sankt Gallen, Switzerland. A geomorphological inventory system applied to geotechnical appraisal of slope stability. *Engineering Geology* 26, 33-68.

Sarkar, S., Kanungo, D.P. and Mehotra, G.S. (1995) Landslide hazard zonation, a case study in Garhwal Himalaya, India. *Mountain Research and Development* 15, 301-309.

Satterlund, D.R. and Haupt, H.F. (1970) Disposition of snow caught by conifer crowns. *Water Resources Research* 6, 649-652.

Schlyter, P., Johnson, P., Nyberg, R., Persson, P., Rapp, A., Jonasson, C. and Rehn, J. (1993) Geomorphic process studies related to climate change in Karkevagge, Northern Sweden – Status of current research. *Geografiska Annaler* 75A, 55-60.

Selby, M.J. (1974) Dominant geomorphic events in landform evolution. *Bulletin of International Association of Engineering Geology* 9, 85-89.

Sidle, R.C. (1986) Groundwater Accretion in Unstable Hillslopes of Coastal Alaska. In: *Conjunctive Water Use. Proceedings of the Budapest Symposium, July 1986.* IAHS Publication No. 156, pp. 335-343.

Sidle, R.C. (1991) A conceptual model of changes in root cohesion in response to vegetation management. *Journal of Environmental Quality* 20, 43-52.

Sidle, R.C. (1992) A theoretical model of the effects of timber harvesting on slope stability. *Water Resources Research* 28, 1897-1910.

Sidle, R.C. and Swanston, D.N. (1982) Analysis of a small debris slide in coastal Alaska. *Canadian Geotechnical Journal* 19, 167-174.

Sidle, R.C. and Tsuboyama, Y. (1992) A comparison of piezometric response in unchanneled hillslope hollows: coastal Alaska and Japan. *Journal of Japan Society of Hydrology and Water Resources* 5, 3-11.

Sidle, R.C. and Wu, W. (1999) Simulating effects of timber harvesting on the temporal and spatial distribution of shallow landslides. *Zeitschrift für Geomorphologie N.F.* 43, 185-201.

Sidle, R.C., Pearce, A.J. and O'Loughlin, C.L. (1985) *Hillslope stability and land use.* Water Resources Monograph, Vol. 11, American Geophysical Union, Washington, DC.

Sidle, R.C., Brown, R.W. and Williams, B.D. (1993) Erosion processes on arid minespoil slopes. *Soil Science Society of America Journal* 57, 1341-1347.

Sidorova, T. (1998) Regime of mudflows and its potential changes due to global warming. *Publication No. 23*, Norwegian Geotechnical Inst., pp. 249-253

Skempton, A.W. and Delory, F.A. (1957) Stability of natural slopes in London clay. In: *Proceedings of 4th International Conference On Soil Mechanics and Foundation Engineering* 4, pp. 379-381.

Starkel, L. (1972) The role of catastrophic rainfall in the shaping of the relief of the lower Himalaya (Darjeeling Hills). *Geographia Polonica* 21, 103-147.

Steiner, F. (1988) Agroforestry's coming of age. *Journal of Soil and Water Conservation* 43, 157-158.

Swanson, F.J. and Dyrness, C.T. (1975) Impact of clearcutting and road construction on soil erosion by landslides in the western Cascades, Oregon. *Geology* 3, 393-396.

Swanson, F.J. and Swanston, D.N. (1977) Complex mass movement terrains in the western Cascade Range, Oregon. *Reviews in Engineering Geology*, Landslides. Vol. 3, Geological Society of America, Boulder, Colorado. pp. 113-124.

Swanston, D.N. (1970) Mechanics of debris avalanching in shallow till soils of southeast Alaska. *USDA Forest Service Research Paper PNW-103*, Portland, Oregon.

Swanston, D.N. (1981) Creep and earthflow erosion from undisturbed and management impacted slopes in the Coast and Cascade Ranges of the Pacific Northwest, U.S.A. *IAHS Publication 132*, pp. 76-94.

Swanston, D.N. and Schuster, R.L. (1989) Long-term landslide hazard mitigation programs: structure and experience from other countries. *Bulletin Association of Engineering Geology* 26, 109-133.

Swanston, D.N. and Swanson, F.J. (1976) Timber harvesting, mass erosion, and steepland forest geomorphology in the Pacific Northwest. In: Coats, D.R. (ed.), *Geomorphology and Engineering*. Dowden, Hutchison and Ross, Stroudsburg, Pa., pp. 199-221.

Temple, P.H. and Rapp, A. (1972) Landslides in the Mgeta area, western Uluguru Mountains, Tanzania. *Geografiska Annular* 54A, 157-193.

Terlien, M.T.J., Van Westen, C.J. and Van Asch, T.W.J. (1995) Deterministic Modelling in GIS-Based Landslide Hazard Assessment. In: Carrara, A. and Guzzetti, F (eds), *Geographical Information Systems in Assessing Natural Hazards*. Kluwer Academic Publishers, Dordrecht, pp. 57-78.

Trenberth, K.E. and Owen, T.W. (1999) Workshop on indicies and indicators for climate extremes, Asheville, NC, USA, 3-6 June 1997, breakout group A: storms. *Climatic Change* 42, 9-21.

Trustrum, N.A., Lambert, M.G. and Thomas, V.J. (1983) The impact of soil slip erosion on hill country pasture production in New Zealand. In: *Proceedings of the 2nd International Conference on Soil Erosion and Conservation, University of Hawaii, Honolulu.*

Tsuboyama, Y., Hosoda, I., Noguchi, S. and Sidle, R.C. (1994) Piezometric response in a zero-order basin, Hitachi Ohta, Japan. In: Ohta, T. (ed.), *Proceedings of International Symposium on Forest Hydrology, University of Tokyo, Japan*. pp. 217-224.

Tsuboyama, Y., Noguchi, S., Shimizu, T., Sidle, R.C. and Hosoda, I. (1998) Intrastorm fluctuations of piezometric head and soil temperature within a steep forested hollow. In: Sassa, K. (ed.), *Environmental Forest Science (Proc. of IUFRO Div. 8 Conf., October 1998, Kyoto Univ., Japan)*, Vol. 54. Forestry Sciences, Kluwer Academic Publisher, Dordrecht, The Netherlands, pp. 475-482.

Tsuboyama, Y., Sidle, R.C., Noguchi, S., Murakami, S. and Shimizu, T. (2000) A zero-order basin – its contribution to catchment hydrology and internal hydrological processes. *Hydroogical Processes* 14, 387-401.

Tsukamoto, Y., Ohta, T. and Noguchi, H. (1982) Hydrological and geomorphological studies of debris slides on forested hillslopes in Japan. *IAHS-AISH Publication 137*, 89-98.

Van Westen, C.J. (1994) GIS in Landslide Hazard Zonation: A Review, with Examples from the Columbian Andes. In: Price, M.F. and Heywood, D.I. (eds), *Mountain Environments and Geographic Information Systems*. Taylor and Francis, London, pp. 135-166.

Varnes, D.J. (1978) Slope Movement Types and Processes. In: Clark, M. (ed.), *Landslide Analysis and Control*. Spec. Rep. 176, Transport. Res. Board, National Academy of Science, National Research Council, Washington, DC, pp. 11-33.

Varnes, D.J. (1984) *Landslide Hazard Zonation: A Review of Principles and Practice*. Natural Hazard No. 3. Commission on the Landslides of the IAEG, UNESCO.

Vorosmarty, C.J., Gutowsky, W.J., Person, M., Chen, T.C. and Chase, D. (1993) Linked atmosphere-hydrology models at the macroscale. In: Wilkinson, W.B. (ed.), *Macroscale Modelling of the Atmosphere. Proceedings of the Yokohama Symposium*. IAHS Publication No. 214, pp. 3-27, Wallingford, UK.

Wang, S.Q. and Unwin, D.J. (1992) Modeling landslide distribution on loess soils in China: an investigation. *International Journal of Geographical Information Systems* 6, 391-405.

Wasson, R.J. and Hall, G. (1982) A long record of mudslide movement at Waerenga-O-Kuri, New Zealand. *Zeitschrift für Geomorphologie N.F.* 26, 73-85.

Watson, A. (1985) Soil erosion and vegetation damage near ski lifts at Cairn Gorm, Scotland. *Biological Conservation* 33, 363-381.

Wells, W.G. (1987) The effects of fire on the generation of debris flows in southern California. In: Costa, J.E. and Wieczorek, G.F. (eds), *Debris flows/avalanches: processes, recognition, mitigation*. Reviews in Engineering geology VII, Geol. Soc. Am., Boulder, Colo., pp. 105-114.

Wieczorek, G.F., Lips, E.W. and Ellen, S.D. (1989) Debris flows and hyperconcentrated floods along the Wasatch Front, Utah, 1983 and 1984. *Bulletin Association of Engineering Geology* 26, 191-208.

Wohl, E.E. and Pearthree, P.P. (1991) Debris flows as geomorphic agents in the Huachuca Mountains of southeastern Arizona. *Geomorphology* 4, 273-292.

Wood, E.F., Lettenmaier, D.P. and Zartarian, G.P. (1992) A land-surface hydrology parameterization with subgrid variability for general circulation models. *Journal of Geophysical Research* 97(D3), 2717-2728.

Wright, C. and Mella, A. (1963) Modifications to the soil pattern of south central Chile resulting from seismic and associated phenomena during the period May to August 1960. *Bulletin of the Seismological Society of America* 53, 1367-1402.

Wu, T.H. and Swanston, D.N. (1980) Risk of landslides in shallow soils and its relation to clearcutting in southeastern Alaska. *Forest Science* 26, 495-510.

Wu, T.H., McKinnel, W.P. and Swanston, D.N. (1979) Strength of tree roots and landslides on Prince of Wales Island, Alaska. *Canadian Geotechnical Journal* 16, 19-33.

Wu, W. and Sidle, R.C. (1995) A distributed slope stability model for steep forested hillslopes. *Water Resources Research* 31, 2097-2110.

Wyss, W. and Yim, S. (1996) Vulnerability and adaptability of Hong Kong to hazards under climatic change conditions. *Water, Air and Soil Pollution* 92, 181-190.

Yin, K.L. and Yan, T.Z. (1988) Statistical prediction model for slope instability of metamorphosed rocks. In: *Proceedings of 5th International Symposium on Landslides, Lausanne*. Vol. 2, pp. 1269-1272.

Yoon, G.G. (1991) Extent and economic significance of landslides in Korea. In: Bell (ed.), *Landslides*. Balkema, Rotterdam, The Netherlands, pp. 1071-1076.

Young, A. (1997) *Agroforestry for Soil Management*. CAB International, Wallingford, UK.

Zhai, P., Sun, A., Ren, F., Liu, X., Gao, B. and Zhang, Q. (1999) Changes of climatic extremes in China. *Climatic Change* 42, 203-218.

Ziemer, R.R. (1981) Roots and the stability of forested slopes. *IAHS Publication No. 132*, 343-357.

Ziemer, R.R. and Swanston, D.N. (1977) Root strength changes after logging in southeast Alaska. *USDA Forest Service Research Note PNW-306*, Portland, Oregon.

Zika, P., Rubar, J. and Kudrna, Z. (1988) Empirical approach to the evaluation of the stability on high slopes. In: *Proceedings of 5th International Symposium on Landslides, Lausanne*. Vol. 2, pp. 1273-1275.

Zimmermann, M. and Haeberli, W. (1992) Climatic change and debris flow activity in high mountain areas - a case study in the Swiss Alps. *Catena Suppliment* 22, 59-72.

Zimmermann, M., Bichsel, M. and Kienholz, H. (1986) Mountain hazard mapping in the Khumbu Himal, Nepal with prototype map, scale 1:50,000. *Mountain Research and Development* 6, 29-40.

Chapter 9

Natural Hazards in Forests: Glacier and Permafrost Effects as Related to Climate Change

Wilfried Haeberli[1] and C.R. Burn[2]
[1]*Department of Geography, University of Zurich, Switzerland*
[2]*Department of Geography and Environmental Studies, Carleton University, Ottawa, Canada*

Atmospheric warming is predicted to be greater in polar regions than at lower latitudes and more pronounced at high altitudes than in lowlands. In polar regions, air and ground warming may lead to a more northerly extension of the boreal forest, as growing seasons lengthen and become warmer. Near-surface permafrost degradation will probably accompany such an evolution in environmental conditions. In some cases slope movement may be catastrophic, but in most instances the settlement is expected to be slow, and the water released by melting ground ice will evaporate. Subpolar forests in permafrost regions are primarily used for firewood and rough lumber, but not construction-grade materials. This is unlikely to change because long-term ground instability, relatively cold soil temperatures, depletion of nutrients in the active layer, and restriction of root systems to the active layer all limit tree growth. Extensive forest fires, usually initiated by lightning after a week or two of hot weather, also deplete timber stocks. There have been suggestions that wildfire may increase following climate warming. Meltwater runoff from glaciated and perennially frozen areas represents only a small portion of the annual water supply, but strongly influences stream flow in lowlands during the warm/dry season. The disappearance of perennial ice above and below the earth surface influences the seasonality of discharge by reducing meltwater production in the warm season and by increasing the permeability of frozen/thawing materials. The latter effect may have strong impacts on soil humidity and growth conditions for forest and tundra in such dry areas as Tian Shan Mountains or Tibet Plateau. In general, accelerated future warming would cause an enlargement of the periglacial belt in high mountain areas, an upslope shifting of hazard processes and a widespread reduction in the stability of formerly glaciated or perennially-frozen slopes. In the case of accelerated future warming, the cryospheric components of high mountain environments would be expected to change at high rates and lead to pronounced disequilibria in the water cycle, in mass-wasting processes and sediment flux, as well as in growth conditions for vegetation. For those directly involved with such changes, the main challenge would be to adapt to the high rates of environment evolution. Empirical knowledge would have to be replaced more and more by improved process understanding, especially concerning runoff formation and slope stability.

Introduction

Earth's snow and ice cover is a critical component of the global climate system. The continental ice sheets of Greenland and Antarctica and the extent of sea-ice actively influence circulation of the atmosphere and the oceans (Budd *et al.*, 1998). The varying spatial extent of glaciers and ice caps through time indicates past changes in surface energy fluxes (Haeberli *et al.*, 1999a), particularly fluctuations in summer temperature and precipitation. In contrast, the duration, thickness and extent of snow cover and freshwater ice provide information about winter conditions. The climatological roles of snow cover are to nourish glaciers, reflect solar radiation reaching the ground surface and regulate ground temperatures in winter.

Permafrost, or perennially-frozen ground, can also contain large amounts of ice. The principal ecological role of permafrost is to provide a near-surface hydrologic barrier that directly influences soil moisture and runoff within the active layer, the ground above permafrost that freezes and thaws each year (French, 1996). Most soil microbial activity and plant root development occurs in the active layer. A potential positive feedback of increasing summer temperature is to alter high-latitude ecosystems from their current status as net sinks for greenhouse gases to sources of methane and carbon as the soil warms (e.g., Weller *et al.*, 1995).

Climate projections indicate that there could be pronounced reductions in the extent of seasonal snow, floating ice and glacier cover during the next century, while the permafrost active layer may deepen considerably. Such changes would probably have significant impacts on related ecosystems; both engineering and agriculture would need to adjust to changes in snow and surface and ground ice conditions (Fitzharris *et al.*, 1996). In order to document potential changes, to assess their impact on ecosystems and socio-economics, and to develop strategies for diminishing potentially-harmful effects or amplifying beneficial ones, global climate observation networks have included cryosphere components in their comprehensive, integrated view of the climate system (Cihlar *et al.*, 1997).

The design of climate-related modelling, observation, mitigation and adaptation strategies involves complex interactions among individual components of the entire climate system. This chapter reviews the main impacts on forest ecosystems of glacier and permafrost evolution during climate change. First, characteristics and the distribution of these two phenomena are outlined. Second, fundamental aspects of their response to climate change are explained. Third, the most prominent effects of such developments on forests in high-mountain and subpolar regions are discussed.

Distribution and Characteristics of Glaciers, Permafrost and Forests in Cold Regions

The distribution of glaciers, permafrost and forests is primarily a function of mean annual air temperature and annual precipitation (Fig. 9.1). In humid-maritime regions, the equilibrium line, separating accumulation from ablation areas on glaciers, is at low altitude, because of the large amount of ablation required to eliminate the deep snowfall (Shumskii, 1964). Temperate glaciers dominate these landscapes. Such ice bodies, with relatively rapid flow, exhibit a high mass

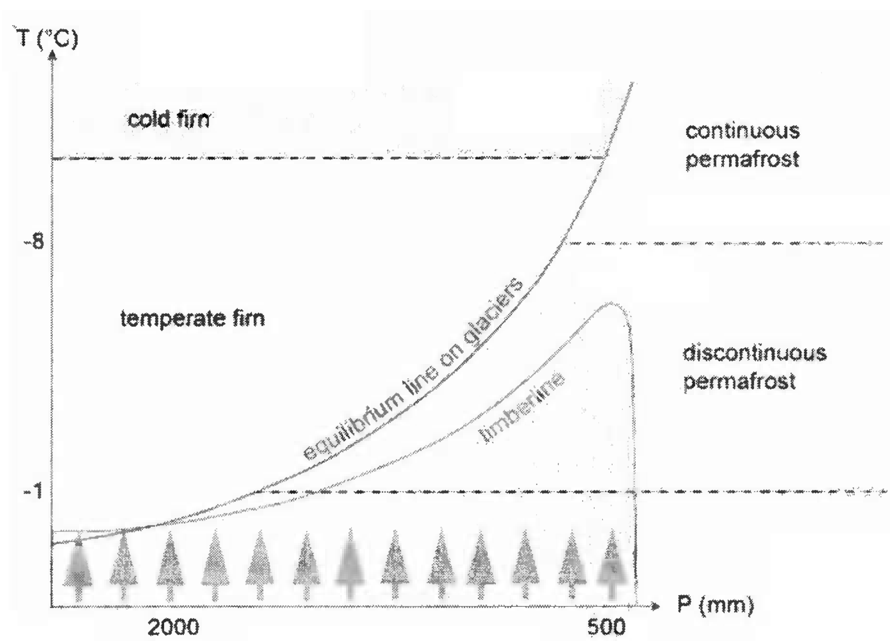


Fig. 9.1. Schematic diagram of glacier, permafrost and forest limits as a function of mean annual air temperature and average annual precipitation. Forests verge on glaciers in humid-maritime climates and grow above permafrost in dry-continental areas (after Shumskii, 1964; Haeberli *et al.*, 1989).

turnover and react strongly to atmospheric warming by enhanced melt and runoff. The ice caps and valley glaciers of Patagonia and Iceland, the western Cordillera of North America and the mountains of New Zealand and Norway are features of this type (Fig. 9.2). The lower parts of such temperate glaciers may extend into forested valleys, where summer warmth and winter snow accumulation prevent development of permafrost. In contrast, under dry-continental conditions, such as in northern Alaska, arctic Canada (Fig. 9.3), subarctic Russia, parts of the Andes near the Atacama desert, and in many central Asian mountain chains, the equilibrium line may be at high elevation. In such regions, polythermal or cold glaciers, lying far above the treeline, have a low mass turnover and are associated with periglacial conditions and permafrost (Shumskii, 1964).

Fig. 9.2. Nigardsbreen, on the east side of Jostedalsbreen ice cap, southern Norway, where precipitation from the Atlantic Ocean is abundant. The glacier descends into permafrost-free, forested terrain. Photograph by W. Haeberli, 1999.

Fig. 9.3. Cold polar glaciers on Axel Heiberg Island, Canadian High Arctic. These glaciers are surrounded by cold, thick permafrost, and lie far from the treeline in an area of polar desert. Photograph by J. Alean, 1977.

An extensive database of the topographic characteristics of glaciers exists in regional glacier inventories (IAHS(ICSI)/UNEP/UNESCO, 1989) and serves as a basis for extrapolating results from observations or modelling of individual glaciers to the regional scale (Oerlemans, 1994). Meier and Bahr (1996) estimated the total number (160,000), area (680,000 km^2), volume (180,000 km^3) and sea-level equivalent (0.5 m) of ice caps and mountain glaciers on Earth and the relative contributions from polar, subpolar, low-latitude/maritime and low-latitude/continental climatic regions (Fig. 9.4). Large areas are covered by ice in the arctic and subarctic mountains of northern Canada and Alaska (275,000 km^2) and the mountains of Eurasia (240,000 km^2), with most of the remaining glaciated area in South America, New Zealand, the subantarctic islands, Iceland and the periphery of Greenland. In general, glaciated area tends to be directly related to distance from densely populated areas, while the economic significance of glaciers is inversely related to this distance. Large maritime glaciers and ice caps provide the main meltwater contribution to present increases in sea level, while meltwater from continental glaciers often influences runoff variability in interior regions of the continents (Kotlyakov and Krenke, 1982).

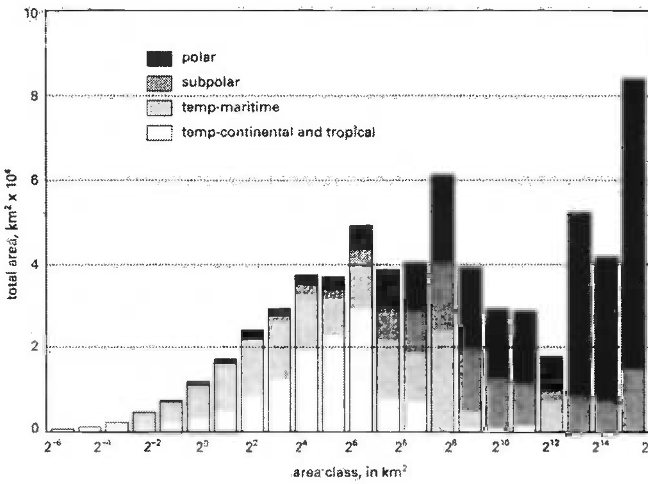


Fig. 9.4. Histogram showing the total global area of glaciers as a function of glacier size. The histogram is divided into four broad climatic regions (modified from Meier and Bahr, 1996).

More than half of Canada and Russia, most of Alaska, a quarter of China and considerable parts of cold mountain ranges worldwide are underlain by permafrost (Brown *et al.*, 1997). The terrain is especially sensitive to disturbance where permafrost is ice-rich, usually in fine-grained sediments, particularly glaciolacustrine and loessal deposits (Brown, 1997a). Coastal and river erosion accelerate in ice-rich ground, while thaw settlement occurs in association with surface disturbance. Ice-rich soil and deeper materials on hill slopes exhibit long-term down-slope movement (Foriero *et al.*, 1998), while deep-seated deformation of ice-supersaturated debris on mountain slopes leads to the development of rock glaciers (Fig. 9.5; Berthling *et al.*, 1998; Hoelzle *et al.*, 1998; Kääb *et al.*, 1998; Haeberli *et al.*, 1999b). The construction of settlements, airfields, pipelines, roads, mines and power dams in such areas requires specialized design techniques. Permafrost in forested areas (Fig. 9.6) is usually discontinuous, underlying less than 80% of the ground. Over large areas of such discontinuous permafrost, the perennially-frozen ground is warmer than -5°C and often warmer than -2°C. Such ground may be particularly sensitive to climate warming which threatens to raise its temperature above 0°C.

Fig. 9.5. Actively creeping ice-rich mountain permafrost ("rock glacier") Suvretta near St Moritz, Swiss Alps. Photograph by W. Haeberli, 1990.

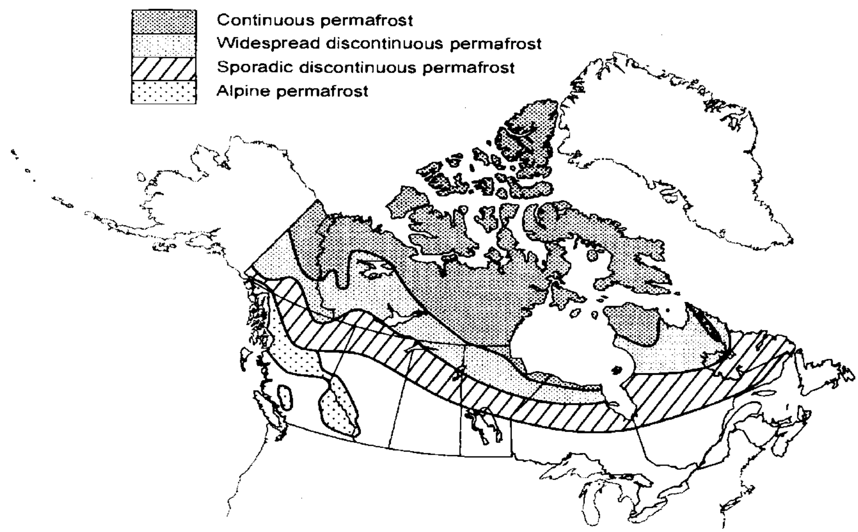


Fig. 9.6. (A) Permafrost map of Canada

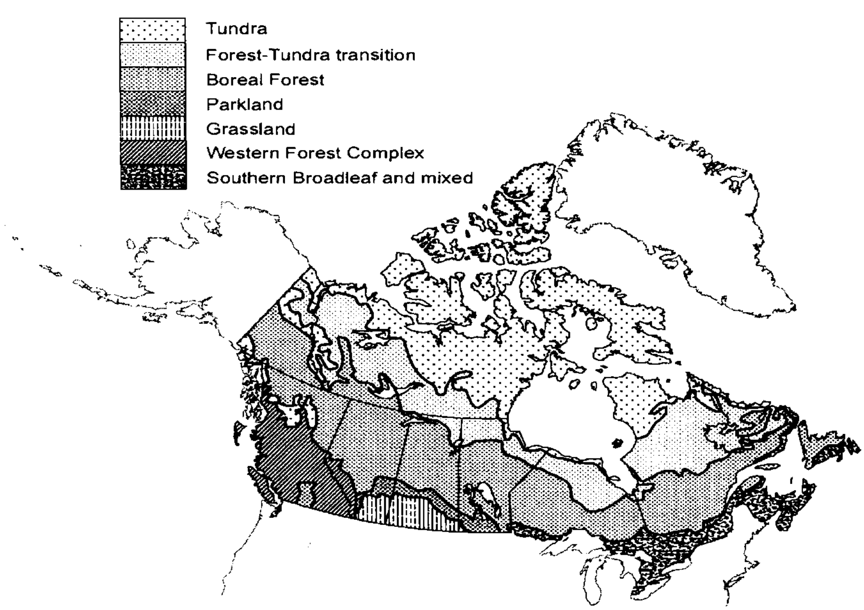


Fig. 9.6. (B) Vegetation map of Canada. Both maps are generalized from the National Atlas of Canada.

Climate Change Effects on Glaciers and Permafrost

Glaciers and discontinuous permafrost may react sensitively to atmospheric warming because of their proximity to the melting temperature. The direct response of glaciers to climate change occurs through changes to the mass balance, and ultimately through variations in glacier length and size (Johannesson *et al.*, 1989). Rates and ranges of such glacier changes can be determined quantitatively over various time intervals and expressed as corresponding energy fluxes. As a result, shifts in glacial cover in mountain areas are key indicators for the early detection of climate change (Haeberli, 1996). Internationally-coordinated long-term monitoring of glaciers started in 1894 and today involves collection and publication of standardized information on the distribution and variability of glaciers in space and time, particularly fluctuations in mass balance and length. Data on special events, such as catastrophic changes, are also available (IAHS (ICSI)/UNEP/UNESCO, 1998).

The response of permafrost to climatic warming occurs through melting at the top of permafrost over periods of years, with or without changes in active-layer thickness. Warming of the temperature profile within permafrost occurs over decades to centuries, and upward displacement of the base of permafrost to reach a new equilibrium thickness takes from centuries to millennia (Osterkamp and Gosink, 1991; Burn, 1998a). In addition to direct climatic effects, principally through changes in air temperature and snowfall, local ground temperature is also strongly influenced by factors related to climate, such as the type of vegetation and properties of the organic layer and soil. These factors can interact in various ways, making it difficult to predict the overall effect of climatic changes (Smith and Riseborough, 1983; Goodwin *et al.*, 1984). Modification of permafrost conditions and subsequent adjustment of related biotic and abiotic surface processes takes place over variable and extended time periods. First attempts are now being made to monitor active-layer thickness and permafrost temperatures within global climate-related observing systems (Cihlar *et al.*, 1997). Efforts presently concentrate on monitoring the thickness of the active layer (Brown, 1997b) and on obtaining borehole temperatures (Isaksen *et al.*, 2000).

Long-term observations of glaciers have provided convincing evidence of rapid climatic change at the global scale; the retreat of mountain glaciers during the 20th century was striking, worldwide. Characteristic average rates of glacier thinning were a few decimetres per year for temperate glaciers and centimetres to a decimetre per year for glaciers in continental areas with firn areas below melting temperature (Meier and Bahr, 1996; Haeberli *et al.*, 1999a). The total retreat of glacier termini is commonly measured in hundreds of metres to a few kilometres, and at retreating glacier termini, the total secular surface lowering is up to several hundred metres (Blair, 1994; Haeberli, 1996; Haeberli *et al.*, 1997). The apparent homogeneity of the signal at the secular timescale, however, contrasts with the large variability at local/regional scales and over time periods of years to decades (Letréguilly and Reynaud, 1990).

Intermittent periods of mass gain and glacier advance are reported from various mountain chains, especially in areas of abundant precipitation such as southern Alaska, New Zealand and Norway (IAHS(ICSI)/UNEP/UNESCO, 1998). Glaciers in the European Alps, on the other hand, have lost ≈ 30 to 40% of their surface area and ≈ 50% of their volume since the mid-19th century, the end of the "Little Ice Age" (Haeberli and Hoelzle, 1995). The recent emergence of a stone-age man from (probably) cold ice on a high-altitude ridge of the Oetztal Alps is a striking illustration which confirms that the extent of Alpine ice is probably less today than during the past 5,000 years (Haeberli *et al.*, 1999a).

Permafrost has also been affected by the recent warming, but its secular evolution is much less understood. Temperatures collected from deep boreholes in Alaska, northern Canada and Europe indicate warming during the past century (Lachenbruch and Marshall, 1986; Taylor, 1991; Isaksen *et al.*, 2000). Permafrost temperatures now appear to be increasing in the European Alps, the Kazakh and Kirghiz Tien Shan, the Tibet Plateau and most of Alaska, though not in the adjacent Yukon Territory (Canada) (Jin *et al.*, 1993; Burn, 1998a; Haeberli *et al.*, 1998; Vonder Mühll *et al.*, 1998; Osterkamp and Romanovsky, 1999).

Glaciated and perennially-frozen regions may be among the most affected if global warming accelerates, but potential changes can only be roughly estimated. Statistical relations and energy balance considerations indicate that one-third to one-half of the current mountain glacier mass could disappear over the next 100 years with a warming of 4°C (Fitzharris *et al.*, 1996; Warrick *et al.*, 1996). With an upward shift of the equilibrium line by some 200 to 300 metres, yearly thickness losses of 1 to 2 metres are expected from temperate glaciers, and many low-latitude mountain chains such as the European Alps would loose major parts of their glacier cover within decades. Large glaciers such as those around the Gulf of Alaska, in Patagonia and in the Himalayas would continue to exist – although greatly reduced – into the 22nd century. Warming of cold firn areas at high altitudes and high latitudes would be pronounced. Various scenarios for the extent of equilibrium permafrost following climate warming indicate that the lower limits of permafrost occurrence in mountain areas could rise by several hundred metres (Fitzharris *et al.*, 1996). However, the time for permafrost to reach new equilibrium thickness may be on the order of centuries, as the elevation of the permafrost base is accomplished by the relatively small geothermal flux (Osterkamp and Gosink, 1991). This time is extended by at least an order-of-magnitude for ice-rich permafrost, where the change in thickness may eventually be tens of metres. Convective effects from groundwater circulation may accelerate rates of permafrost degradation, but these are unpredictable, at present and usually site-specific in scale, although some regional-scale effects have been recognized in northern Alaska (Deming *et al.*, 1992).

Impacts in Low-latitude Mountain Ranges

Glacier and permafrost belts within high-mountain ranges are characterized by extremely steep slopes, large amounts of coarse debris and intense erosion/sedimentation processes. Climate change-related impacts from glaciers and permafrost on forested areas in cold mountain chains can be direct, as in the case of natural hazards to inhabited land, or indirect via changes in landscape evolution, especially with respect to the water cycle and sediment yield. Direct hazards primarily involve the advance or retreat of glaciers, rock falls from destabilized mountain walls, ice avalanches, mudflows from glacier-clad volcanoes, floods from ice- and moraine-dammed lakes, and debris flows from steep permafrost slopes or breaching of moraines (Haeberli *et al.*, 1989, 1997; Major and Newhall, 1989; Evans and Clague, 1993; O'Connor and Costa, 1993; Walder and Costa, 1996; Richardson and Reynolds, 2000).

Glacier advance and retreat

Glacier advances into forested and agricultural land have historically been a major threat in densely-populated mountains (Tufnell, 1984; Grove, 1987). Today, glaciers are predominantly in retreat and the main significance with respect to hazards is the removal of cover from easily eroded moraine slopes, loss of buttresses from rock walls, and the formation of ice- and moraine-dammed lakes. Depending on the availability of quantitative information, predictions of glacier behaviour over a few years must be made by visual inspection in the case of unmeasured glaciers. Using continuity considerations, the behaviour of roughly parameterized glaciers may be estimated over the dynamic response time, typically periods of decades. Predictions may also be made by extrapolation of trends from measured time series. The reliability of such predictions primarily depends on the mass-balance scenario applied (Haeberli *et al.*, 1989). Satellite images have been used to document a glacier surge in connection with the formation of a lake threatening extended areas near Mendoza, Argentina (Espizua and Bengochea, 1990). In regions with maritime climatic conditions, where glacier tongues extend below treeline, recolonization of deglaciated areas by individual trees takes place after a few years, while decades are required to establish closed forests (Burga, 1999). Forest regrowth in deglaciated areas is likely to continue in the future if accelerated atmospheric warming takes place, but lack of soil, unstable slopes and harsh microclimatic conditions may prohibit fast regrowth of vegetation in some places, leaving extended areas of uncovered debris.

Rock falls

Glacier retreat and changes to permafrost may destabilize oversteepened rock walls and accelerate cliff retreat in high-mountain areas. Field and laboratory experiments and analyses of rock glaciers indicate characteristic cliff recession rates on the order of mm per year over the Holocene (Barsch, 1977; Haeberli *et al.*, 1999b). The processes involved, frost shattering and rock fall, have highly variable time and depth scales (Ødegard and Sollid, 1993; Matsuoka *et al.*, 1998). The influence of permafrost thaw on the destabilization of rock walls is thereby a virtually untouched field of research (Dramis *et al.*, 1995). The most important processes concern: (i) fracturing of rocks during seasonal and multi-annual freezing; (ii) changes in hydraulic conductivity, pore-water pressure and circulation during freezing and thawing; and (iii) changes in surface geometry from major rockfalls. Permafrost degradation may cause unfavourable changes in hydraulic conductivity leading to the onset of convective heat transfer and pore-water pressure variations that escape detection as fissures open at depth (Haeberli *et al.*, 1997). Towards the snout of valley glaciers, the pronounced lowering of ice surfaces and vertical loss in valley filling induces a change in the stress field inside the confining mountain walls.

On slopes shaded from direct solar radiation, the lowering of glacier surfaces enables the penetration of permafrost and the formation of ice in rock walls originally covered by temperate ice (Blair, 1994; Wegmann *et al.*, 1998). The penetration of the freezing front into previously unfrozen material has the potential to intensify rock destruction through ice formation in cracks and fissures. Such ice formation, in turn, reduces the near-surface permeability of the rock walls and affects hydraulic pressures inside the fissured rocks. In this way, the general lowering of pore-water pressures in lateral rock walls accompanying the disappearance of temperate glaciers may be counteracted and the rock-wall stability altered. In the absence of permafrost, at low altitudes and on sunny slopes, the stress field within lateral valley walls evolves as water pressure is reduced with decreasing ice thickness. Slope stability may become critical in many instances (McSaveney *et al.*, 1992; Evans and Clague, 1993; McSaveney, 1993; Blair, 1994). The lowest reported values for the overall slope of the run-out path (Evans and Clague, 1988, 1990) indicate that the mobility of rock avalanches in glacial settings may be enhanced by travel on low-friction snow and ice surfaces, generation of basal pore pressure by frictional heating, fluidization through snow- and ice-melt, and channelling or air launching by moraines; forested and even inhabited areas could be affected in the future (Fig. 9.7). While such events remain local in significance, warming is expected to increase the number and frequency of such incidents.

Ice avalanches

A special case of slope instability relates to hanging glaciers. The ablation of many glaciers on steep high-altitude mountain slopes is by ice avalanching, and large blocks of ice may sometimes become detached (Alean, 1985; Dutto *et al.*, 1991). Both glacier and permafrost conditions must be considered when analysing the stability of such steep hanging glaciers. At high altitudes and in the shadow of mountain peaks, firn and ice temperatures are usually well below 0°C. At lower altitudes, however, and on slopes more exposed to the sun, the firn may be at the melting point due to percolating meltwater. In such a situation, only marginal parts of hanging glaciers, especially the vertical front, consist of impermeable, cold ice frozen to the underlying bedrock. This basal temperature pattern probably introduces a thermal anomaly within the underlying permafrost and strong lateral heat flow through the base of the hanging glacier front. The geometry and thermal condition of this ice front, where shear stresses can reach values close to the strength of ice, appears to constitute a key factor controlling the stability of entire ice bodies (Lüthi, 1994). Climate changes may introduce highly complex feedback mechanisms involving surface geometry and firn accumulation, but also temperatures and stress distribution within and at the base of such ice bodies. Long-term monitoring of ice geometries using aerial photography may help to detect unfavourable developments. Independent of attempts to forecast the timing of instability, the run-out distance of potential ice avalanches can be quite realistically assessed (Alean, 1985; Haeberli *et al.*, 1989; Margreth and Funk, 1999). Ice avalanches in connection with earthquake-induced rock avalanches are rare but can be devastating, as in the case of the 1970 mudflow triggered by a rock/ice avalanche at Huascaran, Peru, killing approximately 18,000 people (Körner, 1983; Sidle *et al.*, 1985).

Fig. 9.7. Combined rock/ice avalanche detached from warm permafrost and descending across Brenva Glacier on the southeastern slope of Mt Blanc, 18 January 1998 (c.f. Barla *et al.*, 2000; Deline, 2001).

Ice-covered volcanoes

Lahars and debris flows can originate on active glacier-clad volcanoes. The lateral blast during the eruption of ice-clad Mount St Helens, Washington, on 18 May 1980 destroyed more than 389 km^2 of forest and recreation area, and killed 60 people. The rock avalanche which immediately preceded the eruption, together with intermittent eruptions during the following 3 days, removed nearly 3 km^3 of rocky material from the mountain, including 40 to 45 × 10^6 m^3 of snow and ice (Brugmann and Post, 1981; Lipman and Mullinaux, 1981). Forsyth Glacier, on the northern slope of Mount St Helens, produced two simultaneous ice avalanches that were used to calibrate the timescale of the initial eruption from photographs (Voight, 1981). A series of pyroclastic flows and surges erupting from Nevado del Ruiz, Colombia, on 13 November 1985, mixed with snow and ice on the summit ice cap to trigger catastrophic lahars which killed more than 23,000 people. A total of about 9 × 10^7 m^3 of lahar slurry was transported to depositional areas that extended more than 100 km from the source area (Pierson *et al.*, 1990; Thouret, 1990). Climatic warming has the potential to reduce the ice cover on many active volcanoes and, thereby, decrease corresponding mudflow hazards to forested land.

Glacier floods

Fig. 9.8. Breach in the moraine dam of a glacier, Bhutan Himalaya. Photograph by A. Gansser, 1969.

The largest floods, of up to several km^3, with peak discharges up to several thousand $m^3 s^{-1}$ and lateral reaches of damage up to several tens of km are caused by floods from outbursts of moraine- and ice-dammed lakes or even entire fjords (Blown and Church, 1985; Vuichard and Zimmermann, 1987; Mayo, 1989). Outbursts of large ice-dammed lakes usually take place through flotation of the ice dam followed by progressive enlargement of subglacial channels (Clague and Mathews, 1973; Spring and Hutter, 1981; Clarke, 1982; Evans, 1986), but smaller events may be due to sudden-break mechanisms causing especially high ratios between peak discharge and outburst volume (Haeberli *et al.*, 1989; Walder and Costa, 1996). Breaching of moraine dams (Fig. 9.8) involves piping (progressive groundwater flow) within the morainic material, liquefied flow/slippage on steep slopes, and overtopping with retrogressive incision (Jackson *et al.*, 1989; Haeberli *et al.*, 1997). Glacier shrinkage induced by atmospheric warming may eliminate problems relating to ice-dammed lakes in some places but create new ones in others. Proglacial lakes, growing behind large terminal moraines built up by heavily debris-covered glaciers, however, constitute a recognizable and rapidly-growing hazard, especially in countries such as Nepal and Bhutan (Gansser, 1983; Yamada, 1998).

Debris flows

As a consequence of rapid snowmelt and/or heavy precipitation, debris flows of highly variable size may form in the loose moraines of glacier forefields and at marginal permafrost sites in the scree of debris cones or rock-glacier fronts (Zimmermann and Haeberli, 1992; Rebetez *et al.*, 1997). Thawing of previously frozen material leads to a loss of cohesion, with a simultaneous increase in pore-water pressure in originally ice-rich material and a reduction in resistance to erosion by running water. The transitional conditions of water-saturated fine material remaining on steeply-inclined, thawing permafrost and large caverns originating from the disappearance of massive ground ice bodies lead to extreme hydraulic heterogeneity in non-consolidated materials and are particularly dangerous. Physical and mathematical modelling with respect to debris flow dynamics (e.g., Davis, 1988) has revealed limitations in our present understanding of such important aspects as the transportation of the largest grains at the surface, the depth of erosion along the flow path, or the sudden halting of flow in the runout zone. Empirical approaches to quantifying the most important trigger conditions and flow parameters are discussed by Rickenmann and Zimmermann (1993) based on a large number of Alpine cases. Maximum volume evacuated per unit channel length, for instance, usually remains below 500 to 700 m^3 and the fact that debris flow trajectories have an overall slope greater than about 10° can be used to estimate maximum runout distances (Clague and Evans, 1994). The corresponding hazard to forested and even inhabited land in mountain areas is likely to increase with atmospheric warming but will remain local in significance.

Landscape evolution, water supply and sediments

All of the previously discussed direct hazards have a maximum reach of the order of a few to less than 100 km and are, therefore, of local to regional significance only. The disappearance of large glaciated areas, on the other hand, has the potential to influence the seasonal runoff variability in large rivers which often feed soil and groundwater in semiarid, irrigated regions such as the Ganzu corridor, China, the Argentinian Pampas, and near the Aral Sea. The relative contribution of glacier meltwater during the warm/dry season may be considerable, if not dominant, and disappearance of such summer water as a consequence of widespread deglaciation in the mountainous source areas may lead to shortages detrimental to vast areas of forested and agricultural land (Fitzharris *et al.*, 1996). Changing high-mountain landscapes with intensified erosion of deglaciated and thawed unconsolidated debris could, at the same time, enhance sediment loads in the remaining waters and reinforce accumulation trends in natural and artificial reservoirs as well as in flat riverbeds. In fact, slow trends in large-scale ecosystem functioning could represent stronger impacts than spectacular individual events of low frequency and high magnitude.

Impacts on High-latitude Lowlands

Fig. 9.9. (A) Headwall of a retrogressive thaw slump at Mayo, Yukon Territory, in late September 1986. Ice lenses in ice-rich permafrost are prominent in the lower part of the headwall. The surface vegetation is overhanging, while the distinct boundary between light and dark sediment marks the maximum depth of the active layer in the Holocene.

Fig. 9.9. (B) The ice-rich zone at base of the active layer, shown protruding out from thawing permafrost, July 1984. The ice-rich zone thaws more slowly, due to its higher latent heat content, and then supports the active layer and vegetation overhanging the thawing headwall. Photographs by C.R. Burn.

Permafrost is defined as ground that remains below 0°C for 2 or more years (ACGR, 1988) and, where it occurs, it usually remains frozen for centuries or millennia. At the surface of permafrost terrain there is a seasonally-thawed active layer. The base of the active layer is at the permafrost table, where the maximum temperature during the year is 0°C (Muller, 1947; Burn, 1998b). In unconsolidated sediments, including peat and occasionally in bedrock, the uppermost portion of permafrost (immediately beneath the active layer) is ice-rich (Fig. 9.9; Cheng, 1983; Mackay, 1983; French *et al.*, 1986; Shur, 1988).

This ice-rich ground is the geomorphological unit primarily responsible for the sensitivity of permafrost terrain to disturbance (Mackay, 1970). Ice-rich ground has a water content when thawed that is greater than saturation (Fig. 9.10); the volume above saturation is the excess ice content. Ground with excess ice is sensitive to thawing, but if the ice content is less than or at saturation, the ground is thaw stable.

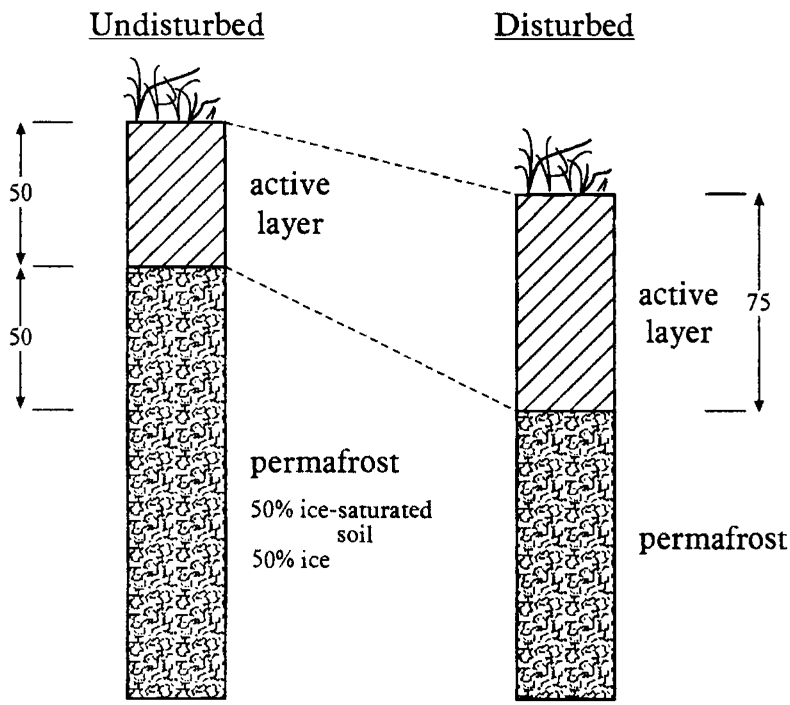


Fig. 9.10. Schematic illustration of ground subsidence with active-layer deepening. In this case, the ice-rich permafrost has an excess ice content of 50%. If the active layer is originally 50 cm thick, then by lowering the base of the active layer a further 50 cm, 25 cm of saturated soil and 25 cm excess ice are melted. Melting of the excess ice leads to 25 cm of surface subsidence (after Mackay, 1970, Fig. 9.3).

The active layer

The thickness of the active layer is a function of the ground surface temperature, the thermal properties of soil materials, and the temperature in permafrost. In general, the active layer is deepest in warm, dry soils or in bedrock, and where the mean annual temperature in permafrost is close to 0°C, while thin active layers are measured in wet peat above cooler ground (Brown, 1970). The active layer generally thickens from north to south through the permafrost regions of the Northern Hemisphere (e.g., Brown, 1970). Beneath the boreal forest of North America, the active layer varies in thickness from about 50 cm to 150 cm (Leverington, 1995; Burn, 1998a); north of the treeline the thickness may decline, but it is rarely less than 30 cm.

Changes in climate that influence active-layer development concern variations in air temperature, snow cover and precipitation. In northern regions, variations in air temperature and precipitation are not independent, most importantly in winter when snowfall is associated with relatively warm conditions. In central Yukon Territory, Canada, warm snowy winters during the 1980s led to permafrost warming by 1° to 1.5°C but subsequently the ground has cooled beneath thin snow covers (Burn, 1998c). A similar trend has been observed in eastern Alaska (Osterkamp and Romanovsky, 1999). On the North Slope of Alaska, active-layer depth increases inland, partly in response to greater trapping of snow by taller vegetation (Romanovsky and Osterkamp, 1995). Indeed, a substantial thinning of the active layer occurs across the treeline associated with the reduction in thickness and increase in density of snow cover at tundra sites, so that cooling of the ground is greater in winter and mean annual ground temperatures are lower (Smith *et al.*, 1998).

Northward migration of forests following summer climate warming may accelerate active-layer deepening through changes to snow-pack properties. This is an example of amplification in climate/permafrost relations due to ecological effects that makes prediction of the impact of climate change on terrain conditions difficult. Field evidence of such amplification has been obtained from examination of active-layer development in western arctic Canada during the early Holocene climate optimum. About 9,000 years ago, when the treeline was considerably further north, and in areas that are now tundra, the active layer was about 2.5 times its present depth (Mackay, 1992). About one half of the active-layer thinning in the last 9,000 years has been attributed to the disappearance of forest from these sites and the remainder to climate cooling (Burn, 1997).

In contrast with the winter regime, warm conditions during summer are usually dry in continental permafrost regions while rain and snowfall are associated with lower temperatures. Evaporation from moist ground also consumes energy and lowers the ground heat flux, further reducing active-layer development. A result is that changes in site wetness may alter ground temperatures and active-layer thickness more than changes in air temperature *per se* (Smith and Riseborough, 1983).

Polar regions are expected to experience greater disturbance than the mid latitudes under enhanced greenhouse warming and changes are expected to be greater in winter than in summer (Roots, 1989). Indeed, records from the last 400 years indicate that the Arctic is warming faster than lower latitudes of the Northern Hemisphere and that the rate of warming was greatest in the 20th century (Overpeck *et al.*, 1997). A wetter climatic regime is projected, with greater snowfall due to a more vigorous general circulation in the atmosphere. For permafrost terrain we should expect a thicker active layer, not primarily because of changes in summer conditions, but cumulatively due to warmer permafrost following deeper snow accumulation.

The most significant terrain hazards wrought by the presence of permafrost are due to melting of the ice-rich zone following deepening of the active layer. Changes in climate may tend to cause relatively gradual deepening of the active layer within the boreal forest because the active layer is relatively thick and climate changes tend to be gradual, superimposed on year-to-year variability. In contrast, surface disturbances, leading to removal of vegetation and organic soil, can cause a substantial, rapid impact on active-layer thickness. The most common and widespread agent of such change is forest fire (Mackay, 1977, 1995; Viereck, 1982). Anthropogenic disturbance, usually from construction or mining, may have a similar effect (Mackay, 1970; Hayley, 1988).

The ice-rich zone at the base of the active layer

The ice-rich zone commonly extends more than a metre below the base of the active layer. Individual ice lenses in this zone are up to several cm thick and ice masses 1 dm thick can occur but are not common. Ice wedges, V-shaped bodies

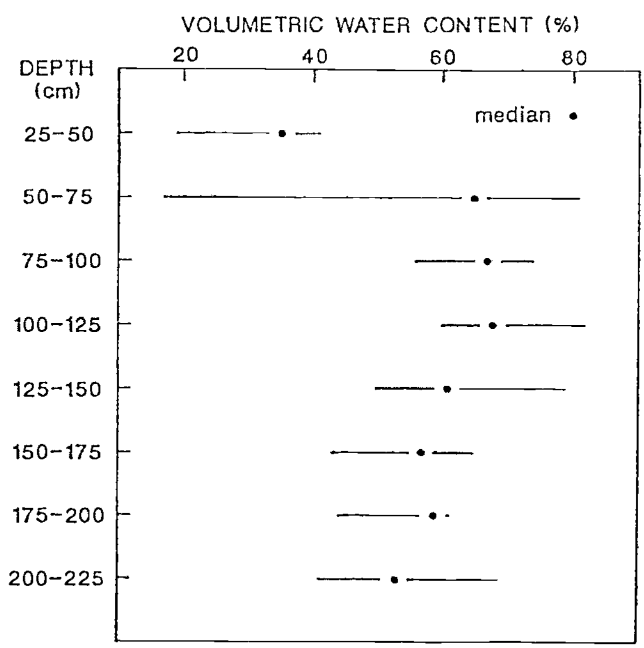


Fig. 9.11. Volumetric water content profile from glaciolacustrine sediments near Mayo, Yukon Territory. The profile shows minimum, median and maximum values of water content in core samples collected from six holes, continuously sampled over 25 cm intervals.

of massive ice, also extend downwards from the base of the active layer (Mackay, 1972; Pollard and French, 1980). Ice wedges are ubiquitous in continuous permafrost, forming polygons 10–15 m in diameter, but they are less evident in discontinuous permafrost, which underlies most of the boreal forest.

A summary profile of the ground-ice content in glaciolacustrine sediments at Mayo, central Yukon Territory, where the dominant vegetation is spruce forest, is presented in Fig. 9.11. The profile indicates that the ice content is characteristically maximal at the top of permafrost. The volumetric water content of these sediments is 40–45% when saturated, so the profile illustrates the high excess ice content in this permafrost.

Data from an experimental disturbance near the sites where the ice contents were derived illustrate ground subsidence following disturbance. Two square plots, 5 m on a side, were established in July 1994, with one left as a control area, while the vegetation and organic mat were removed from the central 9 m^2 of the other. The plots were surveyed over a grid established at 1 m intervals. The survey referenced the ground surface and bottom of the active layer (frost line) in the squares to a nearby benchmark anchored >15 m deep in permafrost. The depth of the active layer was estimated by probing with a pointed steel rod.

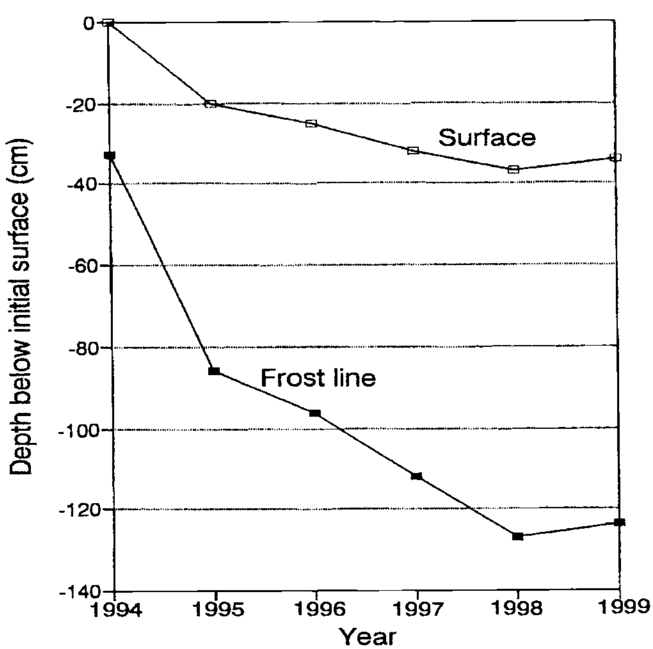


Fig. 9.12. Ground subsidence (1994–1999) at an experimental plot in ice-rich glaciolacustrine sediments near Mayo, Yukon Territory, following surface disturbance. The absolute change in ground surface elevation and the top of frozen ground in late July – early August each year is indicated.

Subsequently the plots have been surveyed each year in late July or August. Fig. 9.12 indicates the lowering of the frost line and the extent of ground subsidence that has occurred from 1994 to 1999 in the disturbed square. At the undisturbed square these variables have shown no trend. The difference between the surface subsidence and the frost line in Fig. 9.12 indicates the depth of the active layer. The rate of subsidence and of active-layer deepening has decreased with time, but the ratio of subsidence to active layer deepening has been about 0.6:1. The precision of the measurements in the irregular microrelief is ±2.5 cm, as determined by repetitive measurements in the undisturbed, control plot. The deeper active layer does not include the thickness of melted ground ice, so the amount of permafrost degradation would have been underestimated if only the active layer had been measured. During five years the ground has subsided by about 35 cm and the active layer has thickened from 33 cm to 90 cm; total permafrost degradation has been 92 cm, with 35 cm from melting of excess ice and 57 cm due to active-layer development.

The original active-layer thickness at the site was 33 cm. For a small thickening of the active layer (e.g., 25%), as may occur during climatic warming, these data indicate ground subsidence of about 5 cm in relatively ice-rich material. In other deposits less ground ice may melt. While these amounts may appear small, such deepening is sufficient to initiate landslides on ice-rich slopes, as in Canada's Mackenzie Valley, following extensive forest fires in summer 1986 (Harry and MacInnes, 1988).

Thermokarst development

Fig. 9.13. Thermokarst lake at Mayo, Yukon Territory, showing trees fallen as a result of lake expansion following thawing of permafrost. Photograph by C.R. Burn, 1988.

The relief produced by melting of ground ice is collectively called thermokarst. Many depressions act as basins for accumulation of water and snow, and once water depth is greater than about one half of the winter ice thickness, permafrost is unsustainable beneath the pool and permafrost degradation proceeds vertically downwards and horizontally (Mackay, 1992). A characteristic of the expansion of thermokarst ponds and lakes in the boreal forest is toppling of trees into the water as permafrost surrounding the lakes thaws (Fig. 9.13). Development of thermokarst lakes requires flat or rolling topography to contain the water-filled depressions.

Fig. 9.14. Aerial photograph of the Mayo permafrost research area, Yukon Territory, showing thermokarst lakes and two retrogressive thaw slumps developed in glaciolacustrine sediments. Part of aerial photograph A27482-84 © 1989 Her Majesty the Queen in Right of Canada, reproduced from the collection of the National Air Photo Library with permission of Natural Resources Canada.

During the Holocene, periods of warm climate have been associated with increased thermokarst development throughout the permafrost regions of Russia (Czudek and Demek, 1970), Alaska (Carson, 1968) and Canada (Harry and MacInnes, 1988; Rampton, 1988). However, individual lakes may be initiated by a site-specific disturbance, such as the melting of an ice wedge, or uprooting of a tree. Changes in surface conditions following forest fire may also lead to ground thawing over a sufficient area for several lakes to develop simultaneously. Although the association of thermokarst development with climate change over millennia is well established, there is no evidence that climatic amelioration since the "Little Ice Age" has led to such development throughout the discontinuous permafrost zones (Fig. 9.6A). Detailed studies at Mayo, Yukon Territory, indicate that site-specific causes are responsible for the development of a cluster of lakes during the last 600 years (Fig. 9.14; Burn and Smith, 1990).

However, at the southerly margin of permafrost in the Northern Hemisphere, there is considerable evidence that permafrost which grew during the "Little Ice Age" is currently melting, as the limits for sustainability of such frozen ground move northwards (Thie, 1974; Halsey *et al.*, 1995). Across the northern Prairie Provinces of Canada recent degradation of permafrost in peatlands is well documented. Such thawing leads to elimination of forests that developed on peat plateaus raised by ice above the surrounding wetland (Vitt *et al.*, 1994). While an association with recent climate warming is proposed for many areas, Zoltai (1993) demonstrated that such changes may also be repeated in a cyclical fashion due to the recurrence of fire at 500-year intervals.

Forest fire

Forest fires cause the most widespread disturbance to surface conditions in permafrost regions. The degree of disturbance is determined by the intensity and nature of the wildfire, which are associated with antecedent meterological conditions. Projections of the nature and frequency of forest fires in a warmer world suggest the extent of damage may increase (Flannigan and Van Wagner 1991). However, certain environments are less susceptible to disturbance by fire, particularly where there is sufficient moisture to reduce the intensity of burning at the surface (Swanson, 1996).

Fires initially raise ground surface temperatures in summer due to reduction in shading (Rouse, 1976). Destruction of vegetation leads to reduced evapotranspiration, raising ground temperatures further as the ground surface dries out. Deepening of the active layer usually follows, so that by the end of summer the ice-rich zone has begun to degrade. On hillslopes, thawing leads to landslides, as the ice-rich zone provides a lubricated plane with low shear strength (Harry and MacInnes, 1988). Such disturbances may also follow particularly warm summers when there is substantial rainfall in early autumn (Mackay and Mathews, 1973), but are almost ubiquitous in sloping ground after fire (Viereck, 1982).

Active-layer deepening is accentuated in years following fire by an increased snow depth in burned areas because interception of snow by the forest canopy is eliminated or reduced. In combination, such changes at the ground surface lead to an increase in ground surface temperature of >2°C and may lead to permafrost eradication. Where permafrost is thin, the degradation may be completed within a decade, but where the thickness is >10 m, centuries are required to thaw the ice-rich ground completely (Burn, 1998a). Vegetation recovery is assured over such timescales, so that permafrost degradation will likely stop. While the most substantial damage occurs in ice-rich terrain, the slowest recovery is at dry sites. Mackay (1977, 1995) carefully documented the active-layer thickening (1968–1993) and recovery following a fire in 1968 at Inuvik, Northwest Territories.

The "drunken forest"

The boreal forest is comprised of a mix of species, but is dominated by white and black spruce (*Picea glauca, P. mariana*), and also contains larch (*Larix laricina*), pine (*Pinus contorta, P. banksiana*), birch (*Betula papyrifera*), poplar and aspen

Fig. 9.15. (A) Deciduous forest of poplar trees (*Populus balsamifera*) growing in permafrost-free ground on a point bar near Mayo, Yukon Territory, July 1991.

Fig. 9.15. (B) A stand of "drunken forest" 9 km north of Mayo, Yukon Territory, July 1994. The trees are black spruce (*Picea mariana*). The ground cover is of lichens and feather mosses. Photographs by C.R. Burn, 1994.

(*Populus balsimifera*, *P. tremuloides*). Inclement conditions during the growing season lead to relatively small trees, rarely >20 m tall, and often <10 m. Where the active layer is relatively deep, usually in dry or permafrost-free ground, trees are usually erect (Fig. 9.15A), but in moist, ice-rich terrain, trees are typically tilted in various directions (Fig. 9.15B). The subsequent woodland is known as a "drunken forest" and is dominated by spruce trees.

Zoltai (1975a) suggested that growth of trees in the boreal forest occurs in cycles punctuated by wildfire. Trees that grow shortly after fire are erect, but as the active layer thins with vegetation succession, tilting occurs. Zoltai (1975a) found that tilted trees were prevalent in stands >100 years old. Growth of ground ice at the base of the active layer may be responsible for the tilting (Mackay, 1983). Separately, Zoltai (1975b) examined the reaction wood in tilted trees to investigate specific conditions causing the tilting. From two locations in the Mackenzie Valley, northwest Canada, he obtained evidence suggesting that tilting is associated with increased autumn temperatures and precipitation, presumably melting the ice-rich zone. This suggestion implies that tilting may become more prevalent following climate change.

Projections of the impact of climate change

The most comprehensive data on the response of permafrost to recent climate change are from Alaska, where ground temperature records are available from a number of sites (Osterkamp and Romanovsky, 1999). Analysis of these data suggests that snow cover plays a critical role on the ground thermal regime because, although air temperatures have not increased in the 1990s, average snow depth has increased and ground temperatures have warmed over large areas. However, in eastern Alaska and in adjacent Yukon Territory, Canada, the inverse has occurred, as snow depths have declined (Burn, 1998c). At present we have no published long-term records of active-layer thickness in the English literature and the analysis of this variable has remained dominantly theoretical. An international effort, the Circumpolar Active Layer Monitoring Program (CALM), has been mounted to obtain such data series (Brown, 1997b).

Nevertheless, Anisimov and Nelson (1996) and Anisimov *et al.* (1997) have taken projections of future climates and attempted to infer the changes they may induce. The equilibrium permafrost distribution in a $2\times CO_2$ world projected by these simulations shows permafrost disappearing from much of its present distribution and the continuous/discontinuous boundary moving several hundred km northward. The time required to reach equilibrium for such transformation is on the order of several millennia.

More useful data have been obtained from geothermal simulations of specific sites, which provide the flexibility to model local ground conditions (Osterkamp and Romanovsky, 1996). These are computationally-intensive, but an important result is that permafrost degradation can often be modelled with few variables (Burn, 1998a). Of these variables, the most critical is the increase in permafrost degradation with the square root of time. This relation implies that for sites with permafrost thickness >10 m degradation will normally take more than a century.

The effects of climatic variability on permafrost stability have also been investigated in the context of climate change. Riseborough and Smith (1993) imposed a variable climate, with similar statistical properties to that of the last 30 years, on the ground thermal regime in warm, thin permafrost characteristic of the boreal forest near Fort Simpson, Northwest Territories. They ran ground temperature simulations for centuries with both stable and warming climate. The work showed that permafrost may become established rapidly due to frost penetration in winter, but the rate of summer thawing is less due to insulation provided by the thawed active layer. In consequence, permafrost has a persistence that is due to the latent heat held in ground ice and the seasonally-changing thermal properties of the surficial layers.

To conclude discussion of permafrost conditions, we reiterate that permafrost is a persistent phenomenon and its future presence is assured within the boreal forest. The terrain hazards presented by permafrost relate to the presence of ground ice. Given that near-surface layers of permafrost terrain are characteristically ice-rich, ground subsidence and slope instability are likely the principal hazards to be expected under climatic change.

Conclusions and Perspectives

Atmospheric warming is predicted to be greater in polar regions than at lower latitudes and more pronounced at high altitudes than in lowlands (Fitzharris *et al.*, 1996; Beniston, 2000). In polar regions, air and ground warming may lead to a more northerly extension of the boreal forest as growing seasons lengthen and become warmer. Increases in snowfall will benefit forest growth by maintaining warmer soil temperatures and increasing soil moisture early in the growing season. Near-surface permafrost degradation is expected to accompany such an evolution in environmental conditions, so that substrate stability is not assured. In some cases slope movement may be catastrophic, but in most instances the settlement may be slow, and the water released by melting ground ice will evaporate. The potentially increasing rate of climatic change, when combined with the slow speed of adjustment in ground-ice conditions and of forest migration over extended distances, is likely to induce long-lasting disequilibria.

Subpolar permafrost regions do not support major commercial forest activities. The forests are used for firewood and rough lumber, but construction-grade materials are produced elsewhere. This is unlikely to change because long-term ground instability, relatively cold soil temperatures, depletion of nutrients in the active layer, and restriction of root systems to the active layer all limit tree growth. Extensive forest fires, usually initiated by lightning after a week or two of hot weather, also deplete timber stocks. There have been suggestions that wildfires may become more common following climate warming. In combination, these effects will probably preclude large-scale forestry operations in permafrost regions, but local enterprises may flourish at sites in the discontinuous zone where there are extensive tracts of unfrozen ground.

In mountain areas, meltwater runoff from glaciated and perennially frozen areas only represents a small contribution to the annual water supply but can have strong influence on streamflow in adjacent lowlands during the warm/dry season. The disappearance of perennial ice above and below the earth surface influences the seasonality of discharge by reducing meltwater production in the warm season and by increasing the permeability of frozen/thawing materials. The latter effect may have strong impacts on soil humidity and growth conditions for forest and tundra in such dry areas as Tian Shan Mountains or Tibet Plateau. Revegetation of terrain following deglaciation is slow under conditions of high-mountain climates and, therefore, deglaciated morainic deposits are exposed to erosion for decades to centuries. In general, accelerated future warming would cause an enlargement of the periglacial belt in high mountain areas, an upslope shifting of hazard processes, and a widespread reduction in the stability of formerly glaciated or perennially-frozen slopes.

Besides densely populated high-mountain areas such as the Alps or parts of the Andes, the described changes would predominantly affect remote areas and have little impact on human settlements. A widespread problem may emerge, however, if the empirical basis for hazard forecasting and protection, i.e., historical documents, statistics of measured time series or traces in nature of past events with long recurrence intervals, lost its applicability as snow and ice conditions evolved beyond the range of Holocene variability. Retreat of glaciers and corresponding changes in landscape and scenery could represent the most directly visible and most easily understood signals of global warming.

In the case of accelerated future warming, the cryospheric components of high mountain environments would likely change at high rates and lead to pronounced disequilibria in the water cycle, in mass-wasting processes and sediment flux, as well as in growth conditions for vegetation. For those directly involved with such changes, the main challenge would be to adapt to the high rates of environment evolution. Empirical knowledge would have to be replaced more and more by improved process understanding, especially concerning runoff formation and slope stability.

Acknowledgements

Scientific activities of WH on glaciers, permafrost and climate change are funded by the University of Zurich, the Swiss National Science Foundation, the EU-PACE project, the Federation of Astronomical and Geophysical Data Analysis Services of ICSU and the International Hydrological Programme of UNESCO. Research by CRB in northwest Canada is supported by the Natural Sciences and Engineering Research Council of Canada; the Polar Continental Shelf Project, Natural Resources Canada; the Inuvik Research Centre of the Aurora Research Institute, Inuvik, Northwest Territories; and the Northern Research Institute, Yukon College, Whitehorse, Yukon Territory. The cartography is by Christine Earl. Polar Continental Shelf Project Contribution #02600.

References

ACGR (Associate Committee on Geotechnical Research) (1988) Glossary of permafrost and related ground ice terms. *National Research Council of Canada, Associate Committee on Geotechnical Research, Technical Memorandum*, 142.

Alean, J.-C. (1985) Ice avalanches: some empirical information on their formation and reach. *Journal of Glaciology* 31, 324-333.

Anisimov, O.A. and Nelson, F.E. (1996) Permafrost distribution in the northern hemisphere under scenarios of climatic change. *Global and Planetary Change* 14, 59-72.

Anisimov, O.A., Shiklomanov, N.I. and Nelson, F.E. (1997) Global warming and active-layer thickness: results from transient general circulation models. *Global and Planetary Change* 15, 61-77.

Barla, G., Dutto, F. and Mortara, G. (2000) Brenva Glacier rock avalanche of 18 January 1997 on the Mount Blanc range, northwest Italy. *Landslide News* 13, 2-5.

Barsch, D. (1977) Nature and importance of mass-wasting by rock glaciers in Alpine permafrost environments. *Earth Surface Processes* 2, 231-245.

Beniston, M. (2000) Environmental change in mountains and uplands. *Key Issues in Environmental Change*. Edward Arnold, London, and Oxford University Press, Oxford.

Berthling, I., Etzelmüller, B., Iken, T. and Sollid, J.L. (1998) The rock glaciers on Prins Karls Forland, Svalbard. I: Internal structure, flow velocity and morphology. *Permafrost and Periglacial Processes* 9, 135-145.

Blair, R.W. (1994) Moraine and valley wall collapse due to rapid deglaciation in Mount Cook National Park, New Zealand. *Mountain Research and Development* 14, 347-358.

Blown, A. and Church, M. (1985) Catastrophic lake drainage within the Homathko River basin, British Columbia. *Canadian Geotechnical Journal* 22, 551-563.

Brown, J. (1997a) Disturbance and recovery of permafrost terrain. In: Crawford, R.M.M. (ed.), *Disturbance and Recovery of Arctic Lands*. Kluwer Academic Publishers, Dordrecht, The Netherlands, pp. 167-178.

Brown, J. (1997b) Circumpolar active layer monitoring (CALM) network – November 1997. *Frozen Ground* 21, 22-23.

Brown, J., Ferrians, O.J.Jr, Heginbottom, J.A. and Melnikov, E.S. (1997) Circum-arctic map of permafrost and ground ice-conditions. U.S. Geological Survey, *Circum-Pacific Map Series*, CP-45.

Brown, R.J.E. (1970) *Permafrost in Canada: its influence on northern development.* University of Toronto Press, Toronto.

Brugmann, M.M. and Post, A. (1981) Effects of volcanism on the glaciers of Mount St. Helens. *US Geological Survey, Circular 850-D.*

Budd, W.F., Coutts, B. and Warner, R.C. (1998) Modelling the Antarctic and Northern Hemisphere ice-sheet changes with global climate through the glacial cycle. *Annals of Glaciology* 27, 153-160.

Burga, C.A. (1999) Vegetation development on the glacier forefield Morteratsch (Switzerland). *Applied Vegetation Science* 2, 17-24.

Burn, C.R. (1997) Cryostratigraphy, paleogeography, and climate change during the early Holocene warm interval, western Arctic coast, Canada. *Canadian Journal of Earth Sciences* 34, 912-925.

Burn, C.R. (1998a) The response (1958-1997) of permafrost and near-surface ground temperatures to forest fire, Takhini River valley, southern Yukon Territory. *Canadian Journal of Earth Sciences* 35, 184-199.

Burn, C.R. (1998b) The active layer: two contrasting definitions. *Permafrost and Periglacial Processes* 9, 411-416.

Burn, C.R. (1998c) Field investigations of permafrost and climatic change in northwest North America. In: Lewkowicz, A.G. and Allard, M. (eds), Proceedings of 7th International Conference on Permafrost, Yellowknife, N.W.T., 23-26 June 1998. Quebec City, Quebec. *Collection Nordicana* 57, 107-120.

Burn, C.R. and Smith, M.W. (1990) Development of thermokarst lakes during the Holocene at sites near Mayo, Yukon Territory. *Permafrost and Periglacial Processes* 1, 161-176.

Carson, C.E. (1968) Radiocarbon dating of lacustrine strands in arctic Alaska. *Arctic* 21, 12-26.

Cheng, G. (1983) The mechanism of repeated-segregation for the formation of thick-layered ground ice. *Cold Regions Science and Technology* 8, 57-66.

Cihlar, J., Barry, T.G., Ortega Gil, E., Haeberli, W., Kuma, K., Landwehr, J.M., Norse, D., Running, S., Scholes, R., Solomon, A.M. and Zhao, S. (1997) *GCOS/GTOS plan for terrestrial climate-related observation.* GCOS 32, version 2.0, WMO/ TD-796, UNEP/DEIA/TR, 97-7.

Clague, J.J. and Evans, S.G. (1994) Formation and failure of natural dams in the Canadian Cordillera. Geological Survey of Canada, *Bulletin 464.*

Clague, J.J. and Mathews, W.H. (1973) The magnitude of jökulhlaups. *Journal of Glaciology* 12, 501-504.

Clarke, G.K.C. (1982) Glacier outburst floods from 'Hazard Lake', Yukon Territory, and the problem of flood magnitude prediction. *Journal of Glaciology* 28, 3-21.

Czudek, T. and Demek, J. (1970) Thermokarst in Siberia and its influence on the development of lowland relief. *Quaternary Research* 1, 103-120.

Davis, T.R.H. (1988) Debris flow surges: a laboratory investigation. *Mitteilungen 96 der VAW/ ETHZ,* 122.

Deline, P. (2001). Recent Brenva rock avalanches (Valley of Aosta): new chapter in an old story? *Geografia Fisica e Dinamica Quaternaria.*

Deming, D., Sass, J.H., Lachenbruch, A.H. and De Rito, R.F. (1992) Heat flow and subsurface temperature as evidence for basin-scale ground-water flow, North Slope of Alaska. *Geological Society of America Bulletin* 104, 528-542.

Dramis, F., Govi, M., Guglielmin, M. and Mortara, G. (1995) Mountain permafrost and slope instability in the Italian Alps. The Val Pola landslide. *Permafrost and Periglacial Processes* 6, 73-81.

Dutto, F., Godone, F. and Mortara, G. (1991) L'écroulement du glacier supérieur de Coolidge (Paroi nord du Mont Viso, Alpes occidentales). *Revue de Géographie Alpine* 2, 7-18.

Espizua, L.E. and Bengochea, J.D. (1990) Surge of Grande del Nevado Glacier (Mendoza, Argentina) in 1984: its evolution through satellite images. *Geografiska Annaler* 72A, 255-259.

Evans, S.G. (1986) The maximum discharge of outburst floods caused by the breaching of man-made and natural dams. *Canadian Geotechnical Journal* 23, 385-387.

Evans, S.G. and Clague, J.J. (1988) Catastrophic rock avalanches in glacial enviroments. Landslides. In: *Proceedings of the Fifth International Symposium on Landslides, 10-15 July 1988, Lausanne.* Vol. 2, pp. 1153-1158.

Evans, S.G. and Clague, J.J. (1990) Reconnaissance observations on the Tim Williams Glacier rock avalanche, near Stewart, British Columbia. *Geological Survey of Canada Paper 90-1E,* 351-354.

Evans, S.G. and Clague, J.J. (1993) Glacier-related hazards and climatic change. *In* The World at Risk: Natural Hazards and Climatic Change. In: Bras, R. (ed.), *American Institute of Physics, Conference Proceedings* 277, 48-60.

Fitzharris, B.B., Allison, I., Braithwaite, R.J., Brown, J., Foehn, P.M.B., Haeberli, W., Higuchi, K., Kotlyakov, V.M., Prowse, T.D., Rinaldi, C.A., Wadhams, P., Woo, M.-K., Xie Youyou, Anisimov, O., Aristarain, A., Assel, R.A., Barry, R.G., Brown, R.D., Dramis, F., Hastenrath, S., Lewkowicz, A.G., Malagnino, E.C., Neale, S., Nelson, F.E., Robinson, D.A., Skvarca, P., Taylor, A.E. and Weidick, A. (1996) The cryosphere: changes and their impacts. Climate Change 1995. Impacts, adaptations and mitigation of climate change: scientific-technical analyses. *Contribution of Working Group II to the Second Assessment Report of the Intergovernmental Panel on Climate Change.* Cambridge University Press, Cambridge, UK, pp. 241-265.

Flannigan, M.D. and Van Wagner, C.E. (1991) Climate change and wild fire in Canada. *Canadian Journal of Forest Research* 21, 66-72.

Foriero, A., Ladanyi, B., Dallimore, S.R., Egginton, P.A. and Nixon, F.M. (1998) Modeling of deep seated hill slope creep in permafrost. *Canadian Geotechnical Journal* 35, 560-578.

French, H.M. (1996) *The periglacial environment*, 2nd edn. Longman, London.

French, H.M., Bennett, L. and Hayley, D.W. (1986) Ground ice conditions near Rea Point and on Sabine Peninsula, eastern Melville Island. *Canadian Journal of Earth Sciences* 23, 1389-1400.

Gansser, A. (1983) Geology of the Bhutan Himalaya. *Denkschriften der Schweizerischen Naturforschenden Gesellschaft 96*, Birkhäuser Verlag, Basel-Boston-Stuttgart.

Goodwin, C.W., Brown, J. and Outcalt, S.I. (1984) Potential responses of permafrost to climatic warming. In: McBeath, J.H. (ed.), *The potential effects of carbon dioxide-induced climatic changes in Alaska*. University of Alaska, Fairbanks, pp. 92-105.

Grove, J.M. (1987) Glacier fluctuations and hazards. *The Geographical Journal* 153, 351-369.

Haeberli, W. (1996) Glacier fluctuations and climate change detection. *Geografia Fisica e Dinamica Quaternaria* 18, 191-199.

Haeberli, W. and Hoelzle, M. (1995) Application of inventory data for estimating characteristics of and regional climate-change effects on mountain glaciers: a pilot study with the European Alps. *Annals of Glaciology* 21, 206-212.

Haeberli, W., Alean, J-C., Müller, P. and Funk, M. (1989) Assessing risks from glacier hazards in high mountain regions: some experiences in the Swiss Alps. *Annals of Glaciology* 13, 96-102.

Haeberli, W., Wegmann, M. and Vonder Muehll, D. (1997) Slope stability problems related to glacier shrinkage and permafrost degradation in the Alps. *Eclogae Geologicae Helvetiae* 90, 407-414.

Haeberli, W., Barsch, D., Brown, J., Guodong, C., Corte, A.E., Dramis, F., Evin, M., Gorbunov, A.P., Guglielmin, M., Harris, C., Harris, S.A., Hoelzle, M., Kääb, A., King, L., Lieb, G.K., Matsuoka, N., Ødegaard, R.S., Shroder, J.F., Sollid, J.L., Trombotto, D., Von der Mühll, D., Shaoling, W. and Xiufeng, Z. (1998) Mapping, modelling and monitoring of mountain permafrost: A review of ongoing programs. *Programme, Abstracts, and Working Group Reports, Seventh International Conference on Permafrost, 23-26 June 1998, Yellowknife, N.W.T.* pp. 233-246.

Haeberli, W., Frauenfelder, R., Hoelzle, M. and Maisch, M. (1999a) On rates and acceleration trends of global glacier mass changes. *Geografiska Annaler* 81A, 585-591.

Haeberli, W., Kääb, A., Wagner, S., Vonder Mühll, D., Geissler, P., Haas, J.N., Glatzel-Mattheier, H. and Wagenbach, D. (1999b) Pollen analysis and ^{14}C age of moss remains in a permafrost core recovered from the active rock glacier Murtèl-Corvatsch, Swiss Alps: geomorphological and glaciological implications. *Journal of Glaciology* 45, 1-8.

Halsey, L.A., Vitt, D.H. and Zoltai, S.C. (1995) Disequilibrium response of permafrost in boreal continental western Canada to climate change. *Climatic Change* 30, 57-73.

Harry, D.G. and MacInnes, K.L. (1988) The effect of forest fires on permafrost terrain stability, Little Chicago-Travaillant lake area, Mackenzie Valley, N.W.T. *Current Research, Part D. Geological Survey of Canada, Paper* 88-1D, pp. 91-94.

Hayley, D.W. (1988) Maintenance of a railway grade over permafrost in Canada. In: *Proceedings of 6th International Conference on Permafrost, 2-5 August 1988, Trondheim, Norway*. Tapir, Trondheim, 3, 43-48.

Hoelzle, M., Wagner, S., Kääb, A. and Vonder Mühll, D. (1998) Surface movement and internal deformation of ice-rock mixtures within rock glaciers at Pontresina-Schafberg, Upper Engadin, Switzerland. In: Proceedings of 7th International Conference on Permafrost, Yellowknife, N.W.T., 23-26 June 1998. Quebec City, Quebec. *Collection Nordicana 57*, 465-471.

IAHS(ICSI)/UNEP/UNESCO (1989) World glacier inventory - status 1988. Haeberli, W., Bösch, H., Scherler, K, Østrem, G. and Wallén, C.C. (eds), Nairobi.

IAHS(ICSI)/UNEP/UNESCO (1998) Fluctuations of Glaciers 1990-1995. Haeberli, W., Hoelzle, M., Suter, S. and Frauenfelder, R. (eds), Zurich.

Isaksen, K., Vonder Muehll, D., Gubler, H.U., Kohl, Th. and Sollid, J.L. (2000) Ground surface temperature reconstruction based on data from a deep borehole in permafrost at Janssonhaugen, Svalbard. *Annals of Glaciology* 31, 287-294.

Jackson, L.E., Hungr, O., Gardner, J.S. and Mackay, C. (1989) Cathedral mountain debris flows, Canada. *Bulletin of the International Association for Engineering Geology* 40, 35-54.

Jin, H., Qui, G., Zhao, L., Wang, S. and Zeng, Z. (1993) Thermal regime of alpine permafrost in the upper reach of Urumqi river, Xinjiang, China. In: *Proceedings of 6th International Conference on Permafrost, 5-9 July 1993, Beijing, China.* Vol. 1, pp. 307-311.

Johannesson, T., Raymond, C.F. and Waddington, E.D. (1989) Time-scale for adjustment of glaciers to changes in mass balance. *Journal of Glaciology* 35, 355-369.

Kääb, A., Gudmundsson, G.H. and Hoelzle, M. (1998) Surface deformation of creeping mountain permafrost. Photogrammetric investigations on Murtèl Rock Glacier, Swiss Alps. In: Proceedings of7th International Conference on Permafrost, Yellowknife, N.W.T., 23-26 June 1998. Quebec City, Quebec. *Collection Nordicana 57*, 531-537.

Körner, H.J. (1983) Zur Mechanik der Bergsturzströme vom Huascaran, Peru. *Hochgebirgsforschung 6*, Universitätsverlag Wagner, Innsbruck, 71-110.

Kotlyakov, V.M. and Krenke, A.N. (1982) Investigations of hydrological conditions of alpine regions by glaciological methods. *IAHS Publication 138*, 31-42.

Lachenbruch, A.H. and Marshall, B.V. (1986) Changing climate: geothermal evidence from permafrost in the Alaskan arctic. *Science* 234, 689-696.

Letréguilly, A. and Reynaud, L. (1990) Space and time distribution of glacier mass balance in the northern hemisphere. *Arctic and Alpine Research* 22, 43-50.

Leverington, D. (1995) A field survey of late-summer depths to frozen ground at two study areas near Mayo, Yukon Territory, Canada. *Permafrost and Periglacial Processes* 6, 373-379.

Lipman, P.W. and Mullinaux, D.R. (1981) The 1980 eruptions of Mount St. Helens, Washington. *US Geological Survey Professional Paper 1250.*

Lüthi, M. (1994) Stabilität steiler Gletscher: Eine Studie über den Einfluss möglicher Klimaänderungen; Untersuchungen am Beispiel eines Hängegletschers in der Westflanke des Eigers. Diploma thesis, ETH Zurich.

Mackay, J.R. (1970) Disturbances to the tundra and forest tundra environment of the western Arctic. *Canadian Geotechnical Journal* 7, 420-432.

Mackay, J.R. (1972) The world of underground ice. *Annals, Association of American Geographers* 62, 1-22.

Mackay, J.R. (1977) Changes in the active layer from 1968 to 1976 as a result of the Inuvik fire. *Report of Activities, part B. Geological Survey of Canada, Paper 77-1B*, 273-275.

Mackay, J.R. (1983) Downward water movement into frozen ground, western Arctic coast, Canada. *Canadian Journal of Earth Sciences* 20, 120-134.

Mackay, J.R. (1992) Lake stability in an ice-rich permafrost environment: examples from the western arctic coast. In: Robarts, R.D. and Bothwell, M.L. (eds), *Aquatic Ecosystems in Semi-arid Regions: Implications for Resource Management.*

Environment Canada, Saskatoon, Sask. National Hydrology Research Institute Symposium Series, 7, 1-26.

Mackay, J.R. (1995) Active layer changes (1968 to 1993) following the forest-tundra fire near Inuvik, N.W.T., Canada. *Arctic and Alpine Research* 27, 323-336.

Mackay, J.R. and Mathews, W.H. (1973) Geomorphology and Quaternary history of the Mackenzie River valley near Fort Good Hope, Northwest Territories, Canada. *Canadian Journal of Earth Sciences* 10, 26-41.

Major, J.J. and Newhall, C.G. (1989) Snow and ice perturbation during historical volcanic eruptions and the formation of lahars and floods. *Bulletin of Volcanology* 52, 1-27.

Margreth, S. and Funk, M. (1999) Hazard mapping for ice and combined snow/ice avalanches - two case studies from the Swiss and Italian Alps. *Cold Regions Science and Technology* 30, 159-173.

Matsuoka, N., Hirakawa, K., Watanabe, T., Haeberli, W. and Keller, F. (1998) The role of diurnal, annual and millennial freeze-thaw cycles in controlling alpine slope instability. In: Proceedings of 7th International Conference on Permafrost, Yellowknife, N.W.T., 23-26 June 1998. Quebec City, Quebec. *Collection Nordicana* 57, 711-717.

Mayo, L. (1989) Advance of Hubbard Glacier and 1986 outburst of Russel Fjord, Alaska, U.S.A. *Annals of Glaciology* 13, 189-194.

McSaveney, M.J. (1993) Rock avalanches of 2 May and 16 September 1992, Mount Fletcher, New Zealand. *Landslide News* 7.

McSaveney, M.J., Chinn, T.J. and Hancox, G.T. (1992) Mount Cook rock avalanche of 14 December 1991. *Landslide News* 6.

Meier, M.F. and Bahr, D.B. (1996) Counting Glaciers: Use of Scaling Methods to Estimate the Number and Size Distribution of the Glaciers on the World. In: Colbeck, S.C. (ed.), *Glaciers, Ice Sheets and Volcanoes: a Tribute to Mark F. Meier*. CRREL Special Report 96-27, US Army, Hanover, New Hampshire, pp. 1-120.

Muller, S.W. (1947) *Permafrost or permanently frozen ground and related engineering problems*. Edwards, Ann Arbor, Michigan.

O'Connor, J.E. and Costa, J.E. (1993) Geologic and hydrologic hazards in glacierized basins in North America resulting from 19th and 20th century global warming. *Natural Hazards* 8, 121-140.

Ødegard, R.S. and Sollid, L.S. (1993) Coastal cliff temperatures related to the potential for cryogenic weathering processes, western Spitsbergen, Svalbard. *Polar Research* 12, 95-106.

Oerlemans, J. (1994) Quantifying global warming from the retreat of glaciers. *Science* 264, 243-245.

Osterkamp, T.E. and Gosink, J.P. (1991) Variations in permafrost thickness in response to changes in paleoclimate. *Journal of Geophysical Research* 96, 4423-4434.

Osterkamp, T.E. and Romanovsky, V.E. (1996) Characteristics of changing permafrost temperatures in the Alaskan Arctic. *Arctic and Alpine Research* 28, 267-273.

Osterkamp, T.E. and Romanovsky, V.E. (1999) Evidence for warming and thawing of discontinuous permafrost in Alaska. *Permafrost and Periglacial Processes* 10, 17-37.

Overpeck, J., Hughen, K., Hardy, D., Bradley, R., Case, R., Douglas, M., Finney, B., Gajewski, K., Jacoby, G., Jennings, A., Lamoureux, S., Lasca, A., MacDonald, G., Moore, J., Retelle, M., Smith, S., Wolfe, A. and Zielinski, G. (1997) Arctic environmental change of the last four centuries. *Science* 278, 1251-1256.

Pierson, T.C., Janda, R.J., Thouret, J.-C. and Borrero, C.A. (1990) Perturbation and melting of snow and ice by the 13 November 1985 eruption of Nevado del Ruiz, Colombia, and consequent mobilization, flow and deposition of lahars. *Journal of Volcanology and Geothermal Research* 41, 17-66.

Pollard, W.H. and French, H.M. (1980) A first approximation of the volume of ground ice, Richards Island, Pleistocene Mackenzie Delta, Northwest Territories, Canada. *Canadian Geotechnical Journal* 17, 509-516.

Rampton, V.N. (1988) Quaternary geology of the Tuktoyaktuk Coastlands, Northwest Territories. *Geological Survey of Canada, Memoir 423.*

Rebetez, M., Lugon, R. and Baeriswyl, P.-A. (1997) Climatic warming and debris flows in high mountain regions: the case study of the Ritigraben, Switzerland. *Climate Change* 36, 371-389.

Richardson, S.D. and Reynolds, J.M. (2000) An overview of glacial hazards in the Himalayas. *Quaternary International* 65/66, 31-47.

Rickenmann, D. and Zimmermann, M. (1993) The 1987 debris flows in Switzerland: documentation and analysis. *Geomorphology* 8, 175-189.

Riseborough, D.W. and Smith, M.W. (1993) Modelling permafrost response to climate change and climate variability. In: *Proceedings of 4th International Symposium on Thermal Engineering and Science for Cold Regions, Hanover, New Hampshire*. US Army Cold Regions Research and Engineering Laboratory Special Report, 93-22: 179-187.

Romanovsky, V.E. and Osterkamp, T.E. (1995) Interannual variations of the thermal regime of the active layer and near-surface permafrost in northern Alaska. *Permafrost and Periglacial Processes* 6, 313-335.

Roots, E.F. (1989) Climate change: high latitude regions. *Climatic Change* 15, 223-253.

Rouse, W.R. (1976) Microclimatic changes accompanying burning in subarctic lichen woodland. *Arctic and Alpine Research* 8, 357-376.

Shumskii, P.A. (1964) *Principles of Structural Glaciology*. Transl. D. Kraus. Dorer, New York.

Shur, Yu.L. (1988) The upper horizon of permafrost soils. In: *Proceedings of 6th International Conference on Permafrost, 2-5 August 1988. Trondheim, Norway*. Vol. 1, pp. 867-871.

Sidle, R.C., Pearce, A.J. and O'Loughlin, C.L. (1985) *Hillslope stability and land use*. Water Resources Monograph, no. 11, American Geophysical Union, Washington, DC.

Smith, C.A.S., Burn, C.R., Tarnocai, C. and Sproule, B. (1998) Air and soil temperature relations along an ecological transect through the permafrost zones of northwestern Canada. In: Proceedings of 7th International Conference on Permafrost, Yellowknife, N.W.T., 23-26 June 1998. Quebec City, Quebec. *Collection Nordicana* 57, 1009-1015.

Smith, M.W. and Riseborough, D.W. (1983) Permafrost sensitivity to climatic change. In: *Proceedings of 4th International Conference on Permafrost, 17-22 July 1983, Fairbanks, Alaska*. National Academy Press, Washington, DC., 1, pp. 1178-1183.

Spring, U. and Hutter, K. (1981) Numerical studies of jökulhlaups. *Cold Regions Science and Technology* 4, 227-244.

Swanson, D.K. (1996) Susceptibility of permafrost soils to deep thaw after forest fires in interior Alaska, U.S.A. and some ecologic implications. *Arctic and Alpine Research* 28, 217-227.

Taylor, A.E. (1991) Holocene paleoenvironmental reconstruction from deep ground temperatures: a comparison with paleoclimate derived from the $\delta^{18}O$ record in an ice core from the Aggassiz Ice Cap, Canadian Arctic Archipelago. *Journal of Glaciology* 37, 209-219.

Thie, J. (1974) Distribution and thawing of permafrost in the southern part of the discontinuous permafrost zone in Manitoba. *Arctic* 27, 189-200.

Thouret, J.-C. (1990) Effects of the November 13, 1985 eruption on the snow pack and ice cap of Nevado del Ruiz volcano, Colombia. *Journal of Volcanology and Geothermal Research* 41, 177-201.

Tufnell, L. (1984) *Glazier Hazards*. Longman, London and New York.

Viereck, L.A. (1982) Effects of fire and firelines on active layer thickness and soil temperatures in interior Alaska. In: *Proceedings, 4th Canadian Permafrost Conference, Calgary, Alta, 2-6 March 1980*. National Research Council of Canada, Ottawa, Ontario, pp. 123-135.

Vitt, D.H., Halsey, L.A. and Zoltai, S.C. (1994) The bog landforms of continental western Canada in relation to climate and permafrost patterns. *Arctic and Alpine Research* 26, 1-13.

Voight, B. (1981) The 1980 eruptions of Mount St. Helens, Washington. Time scale for the first moments of the May 18 eruption. *US Geological Survey Professional Paper 1250*, 69-86.

Vonder Mühll, D., Stucki, T. and Haeberli, W. (1998) Borehole temperatures in alpine permafrost: a ten year series. In: Proceedings of 7th International Conference on Permafrost, 23-26 June 1998,Yellowknife, N.W.T. Quebec City, Quebec. *Collection Nordicana* 57, 1089-1095.

Vuichard, D. and Zimmermann, M. (1987) The 1985 catastrophic drainage of a moraine-dammed lake Khumbu Himal, Nepal: cause and consequences. *Mountain Research and Develeopment* 7, 91-110.

Walder, J.S. and Costa, J.E. (1996) Outburst floods from glacier-dammed lakes: the effect of mode of laid drainage on flood magnitude. *Earth Surface Processes and landforms* 21, 701-723.

Warrick, R.A., Le Provost, C., Meier, M.F., Oerlemans, J., Woodworth, P.L., Alley, R.B., Bindschadler, R.A., Bentley, C.R., Braithwaite, R.J., de Wolde, J.R., Douglas, B.C., Dyurgerov, M., Flemming, N.C., Genthon, C., Gornitz, V., Gregory, J., Haeberli, W., Huybrechts, P., Jóhannesson, T., Mikolajewicz, U., Raper, S.C.B., Sahagian, D.L., van de Wal, R.S.W. and Wigley, T.M.L. (1996) Changes in sea level. Climate Change 1995: The Science of Climate Change. *Contribution of Working Group I to the Second Assessment Report of the Intergovernmental Panel on Climate Change.* Cambridge University Press, UK, pp. 359-405.

Wegmann, M., Gudmundsson, G.H. and Haeberli, W. (1998) Permafrost Changes in Rock Walls and the Retreat of Alpine Glaciers: a Thermal Modelling Approach. *Permafrost and Periglacial Processes* 9, 23-33.

Weller, G., Chapin, F.S., Everett, K.R., Hobbie, J.E., Kane, D., Oechel, W.C., Ping, C.L., Reeburgh, W.S., Walker, D. and Walsh, J. (1995) The Arctic flux study: a regional view of trace gas release. *Journal of Biogeography* 22, 365-374.

Yamada, T. (1998) *Glacier lake and its outburst flood in the Nepal Himalaya.* Data Center for Glacier Research, Japanese Society of Snow and Ice, Monograph 1, Japan.

Zimmermann, M. and Haeberli, W. (1992) Climatic change and debris flow activity in high mountain areas; a case study in the Swiss Alps. *Catena Supplement* 22, 59-72.

Zoltai, S.C. (1975a) Structure of subarctic forests on hummocky permafrost terrain in northwest Canada. *Canadian Journal of Forest Research* 5, 1-9.

Zoltai, S.C. (1975b) Tree ring record of soil movements on permafrost. *Arctic and Alpine Research* 7, 331-340.

Zoltai, S.C. (1993) Cyclic development of permafrost in the peatlands of northwestern Alberta, Canada. *Arctic and Alpine Research* 25, 240-246.

Chapter 10

Global Changes, Mangrove Forests and Implications for Hazards along Continental Shorelines

David Taylor[1] and Peta G. Sanderson[2]
[1]*Department of Geography, Trinity College, University of Dublin, Dublin, Republic of Ireland*
[2]*University of Notre Dame, Australia, School of Environmental Management, Freemantle, Western Australia*

Coastal zones are becoming more hazardous due to increased population and greater susceptibility to flooding and salinization, as a result of compaction and subsidence, surface sealing and loss of buffer zones between the coast and settled areas. These conditions are exacerbated in the developing (largely equatorial) world, where rates of population increase are highest and protection schemes and mitigation strategies among the most poorly advanced. The major climate change threat to coastal forest environments is the projected increase in sea level and associated vulnerability of coastal land and inhabitants. In many places, existing coastal forests will be squeezed between rising sea levels and agricultural land, while farmers will have to contend with higher instances of damaging floods and increased salinity. The absence of adaptation strategies, such as the adoption of salt-tolerant crops and associated technologies, will force the abandonment of formerly productive land. Displacement of people will increase population pressures inland. The prognosis for forested coastlines in equatorial regions is poor and is unlikely to improve while coastal environments remain locations for the juxtaposition of dynamic natural environments with relatively static, ecologically disruptive, economic activities. Mangrove forests face the combined threat of global change-induced variations in environmental conditions and clearance and over-exploitation by humans. By degrading mangrove forests, however, humans will remove one of the few effective means of protecting coastal areas from the impacts of global change. Where removal of coastal forests proceeds

along with the construction of new drainage channels and the canalization of existing rivers, impacts of flood surges and intrusions of saline water are likely to be felt sooner and over larger areas than on undisturbed coastal plains. Human activity may therefore enhance the possibility of catastrophic events in the future and, therefore, their own vulnerability to cumulative hazards.

Introduction

Coastlines support a variety of ecosystems and socio-economic activities. They are, however, considered particularly vulnerable to global changes (IPCC, 1997; Klein and Nicholls, 1999; Nicholls *et al.*, 1999). Relatively slight variations in sea level, for example, may have profound impacts over considerable horizontal distances on gently sloping, coastal plains. Moreover, coastal areas of all types will be affected by global warming-induced changes to storm surges and by alterations in the extent, form and intensity of human activity. Humans are already having major direct impacts on coastal areas, with around one fifth of the world's population living within 30 km of the coast (Gommes *et al.*, 1997), through the conversion of coastal ecosystems to agricultural and urban land and the development of infrastructural and coastal protection schemes. These anthropogenic and demographic changes contribute to greenhouse gases that alter climates and sea levels (Hulme *et al.*, 1999a) and modify the hydrology of coastal areas through the imposition of dams, drainage schemes and extensive surface sealing.

Hazards can be viewed as one outcome of the interaction between humans and their environment (Burton *et al.*, 1993). Thus, the presence of humans has the potential to translate a sensitivity of coastlines to global changes into a vulnerability of people inhabiting coastal environments to hazards (Gares *et al.*, 1994). By 1990, around 200 million people were living along hazardous coastlines around the world, defined as beneath the 1 in 1000 year storm surge elevation on the coastal plain (Hoozemans and Hulsbergen, 1995). Rates of population increase in coastal areas are thought to be currently twice the global average (Bijlsma *et al.*, 1996), while increases in sea level, storm intensities and the frequency of flooding are predicted for this century (Wigley, 1995; Saunders, 1999). Coastal areas and the people who live and work there are therefore expected to become increasingly vulnerable to hazards in the coming decades. Some of this increased vulnerability will be transferred to inland areas, as a result of saltwater intrusion along drainage networks and enforced migrations of people from inundated land.

The response of people to a threatened hazard is determined by the perception of the hazard, or complex of hazards, and by opportunities to adapt and adjust lifestyles and livelihoods. Experience, culture and access to information influence the perception of a hazard, whereas opportunities to respond are affected by factors such as age, socio-economic conditions and government policies. Thus, not only are the nature and environmental impacts of future global changes difficult to predict, it is virtually impossible to anticipate – with a high level of confidence – the ways in which humans are likely to respond.

This chapter describes the implications for populated, forested coastlines in equatorial regions of widely accepted, global change scenarios. In particular, the chapter focuses on mangrove-dominated coastlines and estuaries, as mangroves are by far the most prominent coastal forests throughout the tropics. Mangrove forests provide consumptive and protective functions on populated coastlines. They are important sources of natural resources and, although empirical studies are few, computer models have demonstrated the effectiveness of mangrove forests in reducing the impact of storm surges (Massel *et al.*, 1999). Scenarios of global change referred to in this chapter are based upon output from two ocean-atmosphere coupled Global Circulation Models (GCMs). These are the HadCM2 and HadCM3 GCMs that have been developed at the Hadley Centre of the UK's Meteorological Office. HadCM2 output had a major influence on the IPCC's Second Assessment Report (IPCC, 1995); HadCM3 also played a similar role in the IPCC's Third Assessment Report, released in 2001.

Forested Equatorial Coastlines

Whitmore (1998) lists three formations of coastal forest in equatorial regions – mangrove forest, brackish water forest and beach vegetation. Of these, mangrove forest is the most relevant to a discussion that focuses on hazard. Mangrove ecosystems are relatively extensive; about 25 % of coastlines in the tropics are fringed by mangroves (Woodroffe, 1995). Their importance is generally acknowledged as a source of many resources, either directly (e.g., timber) or indirectly (e.g., by serving as feeding and spawning grounds for many species of fish that are harvested commercially). Mangroves also act as sediment traps and provide some protection to ecosystems and human activities inland of the coastal fringe, by acting as natural breakwaters. As a result, the livelihoods of significant numbers of people, including farmers who may have little direct contact with the sea, are in some way dependent upon the presence of intact mangrove forest.

Mangrove forest also occurs on the seaward edge of brackish water forest. This is a relatively species-poor forest type that, in Southeast Asia, is dominated by the palm *Nypa fruticans*. The palm *Raphia taedigera* occupies a similar niche in Central and South America, and may form monospecific stands (Phillips *et al.*, 1997). Brackish water forest extends along the lower, more saline parts of estuaries and grades into freshwater swamp and dryland forests where the opportunity exists. By comparison, beach forest, being confined to sandy ridges that have borne the brunt of much tourist development in recent years, is now limited in extent. Species typical of beach forest in Southeast Asia include *Ardisia elliptica*, *Barringtonia asiatica*, *Cocos nucifera*, *Hibiscus tiliaceus* and *Pandanus bidas* (Whitmore, 1984).

Distribution and biogeography of mangrove forest

The majority of plants that comprise mangrove forest are shrubs and trees, with members of the families *Avicenniaceae*, *Combretaceae*, *Plumbaginaceae*, *Rhizophoraceae* and *Sonneratiaceae* predominating. Many of the plants share a similar physiognomy, including evergreen habit; waxy, salt excreting leaves; negatively geotropic aerating roots and vivipary. On a global scale, the distribution limits of mangrove taxa in the northern and southern hemispheres, respectively 32°N and 37°S, are determined by the incidence of frost and by the availability of suitable physiographic conditions (Chapman, 1976). Duke (1992; p. 70) notes that a close dependence on temperature is apparent: the latitudinal distribution of mangrove taxa is more restricted on eastern continental margins than western, because of the effect of cold ocean currents, and generally matches the 20°C winter isotherm. There are three exceptions, however, all in the southern hemisphere and Duke (1992) suggests these might be relicts of more poleward distributions in the past (Fig. 10.1).

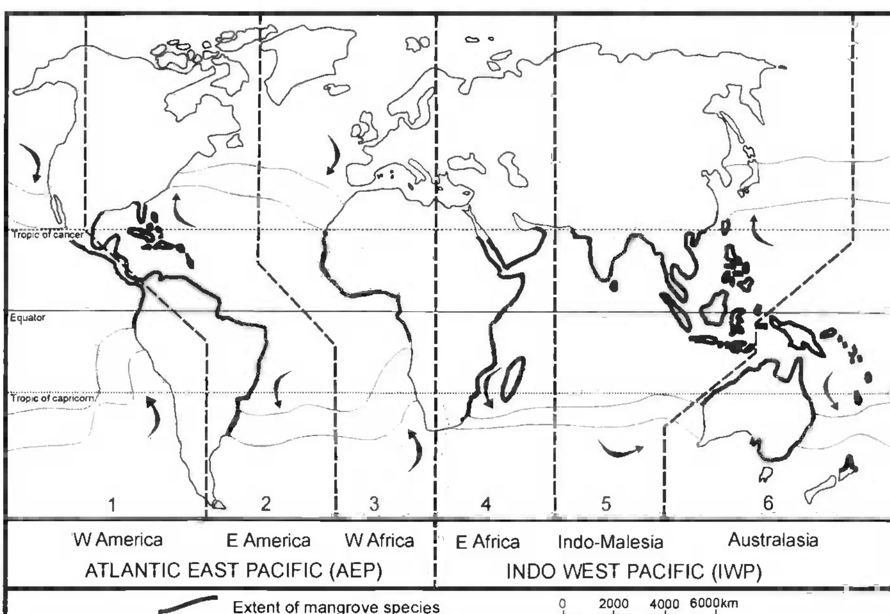


Fig. 10.1. Bioclimatic global distribution limits (shaded areas) of mangrove taxa (from Duke, 1992).

Mangrove taxa can be divided into two biogeographic regions, the Atlantic East Pacific (or New World) and the Indo West Pacific (or Old World). This division is believed to represent the proficiency of the continental landmass of Africa and the Pacific Ocean as barriers to dispersal. Greatest mangrove diversity is found in the Indo West Pacific region (around six times more species than in the Atlantic East Pacific, according to Chapman (1976)) and in particular Southeast Asia, where 36 of the world total of around 60 mangrove taxa have been recorded (Knox, 1986). Further division of these two major biogeographic regions into six sub-regions is possible, based on intra-regional differences in distribution patterns.

On a local scale, mangrove taxa colonize a range of substrates between the highest level of spring tide and mean sea level in low energy coastal and estuarine environments. Mangrove ecosystems develop most fully on gently sloping coastal plains where broad intertidal zones along with abundant supplies of fine-grained sediment and freshwater prevail (Walsh, 1974; Clough, 1992; Woodroffe, 1992). Mangrove forests also require protection from prevailing strong winds and wind-generated waves (Woodroffe, 1995). They are therefore most extensive on sheltered coastlines or on shorelines that are aligned parallel to the prevailing winds.

Thom (1982) described five distinct terrestrial settings, in which mangrove forest is commonly found, based upon a classification of deltas proposed by Wright *et al.* (1974). Contrary to Thom's (1982) classification, many areas of mangrove forest are actually influenced by both rivers and tides and therefore occupy the ecotone between freshwater, saline and terrestrial habitats. Because of this, Woodroffe (1992, 1995) proposes an alternative classification of mangrove ecosystems that integrates environmental setting with some indication of the predominant geomorphological, hydrological and sedimentological influences. Thus, mangrove ecosystems are divided into three forms (Fig. 10.2). First, river-dominated, where salinity is generally low, movement of water is predominantly from the land to the sea and sediments are largely terrigenous. Such mangrove ecosystems may supply considerable amounts of organic material to the coastal zone. Second, fringe or tide-dominated mangroves, in which bidirectional tidal flows and the resuspension of sediments are predominant. Third, interior or wave-dominated mangroves, have limited – or in some cases no – marine connection in terms of water flow. These may act as major sinks for sediments and accumulate peat.

Variations in the frequency of inundation by the sea and the degree of freshwater input can lead to a zoning of mangrove taxa. The lowest point of the shoreline, where inundation by the sea is most frequent, is frequently colonized by a small subset of tree species, comprising *Sonneratia alba*, *Avicennia* spp. and *Rhizophora* spp. Farther back from the shoreline species diversity is increased. The *Rhizophoraceae* family, including *Rhizophora* spp. and *Bruguiera* spp., often dominate this middle zone. *Acanthus* spp. and *Avicennia* spp. are common components on the landward ecotone of mangrove, which generally occupies those parts of the coastal plain only flooded by exceptional tides. Where conditions permit, the inland margins of mangroves are colonized by the palms *Nypa fruticans* and *Phoenix paludosa*, and grade eventually into freshwater swamp forest.

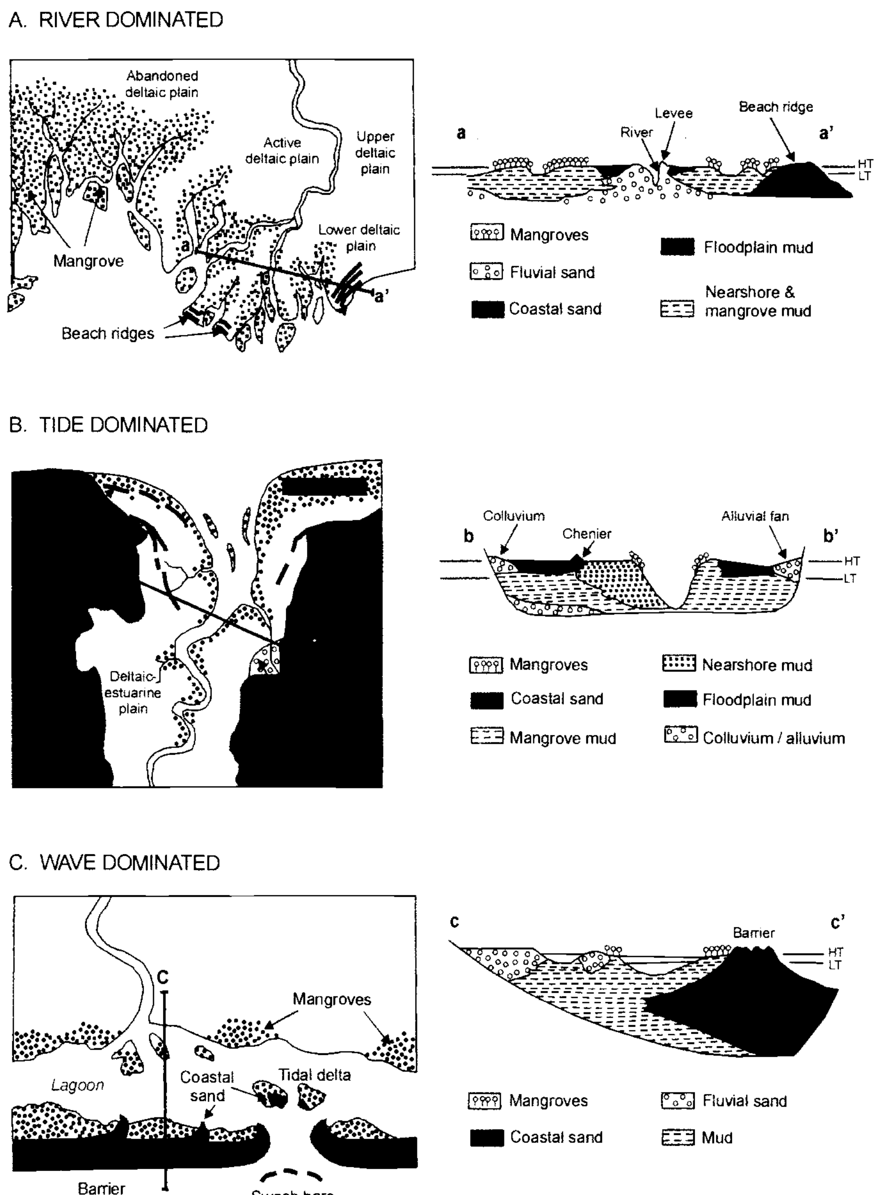


Fig. 10.2. Classification of mangrove forest according to environmental setting. A) River-dominated, based on the Purari delta, Papua New Guinea; B) Tide-dominated, based on macrotidal estuaries in northern Australia; C) Wave-dominated. Illustrations are typical planform (left) and stratigraphic cross-section (right) (based on Woodroffe, 1992).

Coastal Dynamics

Coastal and estuarine processes in equatorial regions are highly dynamic and this variability has great ecological significance. Dynamism over short timescales is largely diurnal and driven by tidal variations. Seasonal perturbations also occur, however, particularly in levels of freshwater input in those areas influenced by monsoonal airflows. Other ecologically important factors that vary over relatively short timescales are the timing, frequency and severity of temperature extremes, lightning and storm surges.

Blasco *et al.* (1996) argue that mangrove taxa are stenotypic to a degree that even minor variations in environmental conditions can result in noticeable mortality. Furthermore, the productivity of mangrove forest is influenced by short-term variations in salinity (Clough, 1992; Snedaker, 1995), whereas frost, lightning and storms influence their composition and structure (Lugo and Patterson-Zucca, 1977; Smith and Duke, 1987; Smith, 1992). Strong wind and wave activity, generated by cyclones, can alter the location and shape of offshore bars and seagrass vegetation (Short and Neckles, 1999) and therefore the exposure of mangroves to subsequent storm surges. Significant damage to coastal vegetation is observed after cyclones, particularly in those areas where their occurrence is infrequent and where the vegetation is shallow-rooted. For example, Woodroffe and Grime (1999) indicate that the present zoning of mangrove taxa and associated substrates in Shoal Bay, northern Australia, is partly a reflection of the devastation caused by Cyclone Tracey, which hit the area in late 1974. Moreover, the high levels of rainfall that are associated with some storms can dramatically alter the supply of sediments and their distribution within the coastal zone (Snedaker, 1995).

Variations in climate, geomorphology and sea level force changes to coastal areas over long timescales. The effects of these are difficult to separate and in many cases interlinked. Thus, changes in global climates affect eustatic sea levels and the intensity of cyclonic activity, and therefore the impact of storm surges. Similarly, some geomorphologic processes may have impacts that are virtually identical to those of sea-level rises. For example, processes such as erosion and tectonism can result in marine inundation. Subsidence in coastal areas, and therefore increased threat of inundation, may also be due to the over-exploitation of aquifers, as appears to be the case for Bangkok, Jakarta and Manila (Nicholls, 1995). Levels of impact of the various forcing factors are a function of the direction, rate and magnitude of change as well as antecedent conditions, and will be modified by local environmental conditions, such as the availability of sediment and the frequency and intensity of storms. As a result of this dependence on initial conditions and non-linearity, coastlines and the ecosystems they support can be viewed as chaotic systems. One consequence of systems exhibiting chaotic behaviour is that their evolution through time is theoretically deterministic but practically – given current levels of knowledge and understanding – not, at least to a high level of certainty. A consequence of non-linearity is that a relatively small perturbation can cause a disproportionate – even catastrophic – response (Scheidegger, 1994).

Mangroves and sea level

Of the environmental perturbations affecting mangroves in equatorial regions, sea level variations have probably had the greatest long-term impact (Woodroffe and Grindrod, 1991). Anderson (1964, 1983) and Anderson and Muller (1975) argue that mangrove taxa are important in the stabilization of alluvium during marine regressions. As the coastline progrades, the colonizing mangrove community gives way to progressively less salt-tolerant taxa and – in some cases – peat-forming conditions. More recently, Furukawa and Sabiham (1985) and Furukawa (1994) have stressed the weaknesses in this model, the major ones being its failure to accommodate the response of mangroves to marine transgressions and the ecological impact of variations in non-marine influences, such as inputs of freshwater and terrigenous sediments (Snedaker, 1995).

It is generally assumed that the mass of mangrove roots trap silt that, together with organic material from the plants themselves, leads to an accumulation of sediment. According to Ellison and Stoddart (1991) and Ellison (1993), an apparent ability to enhance siltation allows mangroves generally to keep pace with a sea level rise of up to approximately 0.08 to 0.09 cm year^{-1}, while higher rates place mangroves under severe ecological stress. The same authors go on to suggest that 'collapse' of mangrove ecosystems is possible at rates of sea level rise in excess of 0.12 cm year^{-1}. Field (1995) contends that these estimates are too low and that mangroves can accumulate sediments at a far higher rate where there are plentiful supplies of fresh water and silt. High rates of sea level rise are likely to force a lateral displacement of mangrove forest or, where such a response is impossible, some destruction of the habitat. However, under certain conditions, such as in deltaic areas where influxes of sediment are large, it is possible to envisage a locally accreting coastline and a seaward expansion of mangrove during a period of rising sea levels along other parts of the same shoreline.

We know that mangrove forests have coped with substantial shifts in sea level in the past, from studies of evidence preserved in coastal sediments of Quaternary age. These shifts have been far greater in absolute terms than those predicted for this century, and at times have been rapid (Williams *et al.*, 1998). Generally, eustatic sea levels rose rapidly once deglaciation commenced towards the end of the last ice age. Relative rates of rise in excess of 0.3 cm year^{-1} are recorded by geomorphic and sedimentary evidence for some sites (Williams *et al.*, 1998). Deglaciation and sea level change do not appear to have progressed smoothly, with at least one major oscillation occurring close to the transition between the last ice age and the present interglacial (the Holocene). Although there is a paucity of well-dated evidence and problems over its interpretation (Woodroffe, 1988, 1990), local variations in sea level rise during the Holocene are reported for equatorial regions (Fig. 10.3). Thus, different generalized sea level curves have been constructed for islands in the tropical Atlantic and Pacific oceans and for continental margins. Glacio-isostatic adjustments will have played a relatively minor role in sea level histories in the tropics. Of greater importance were tectonic activity, subsidence and hydro-isostacy, the latter caused by differences in the level of thermal expansion and the loading of water on the seabed.

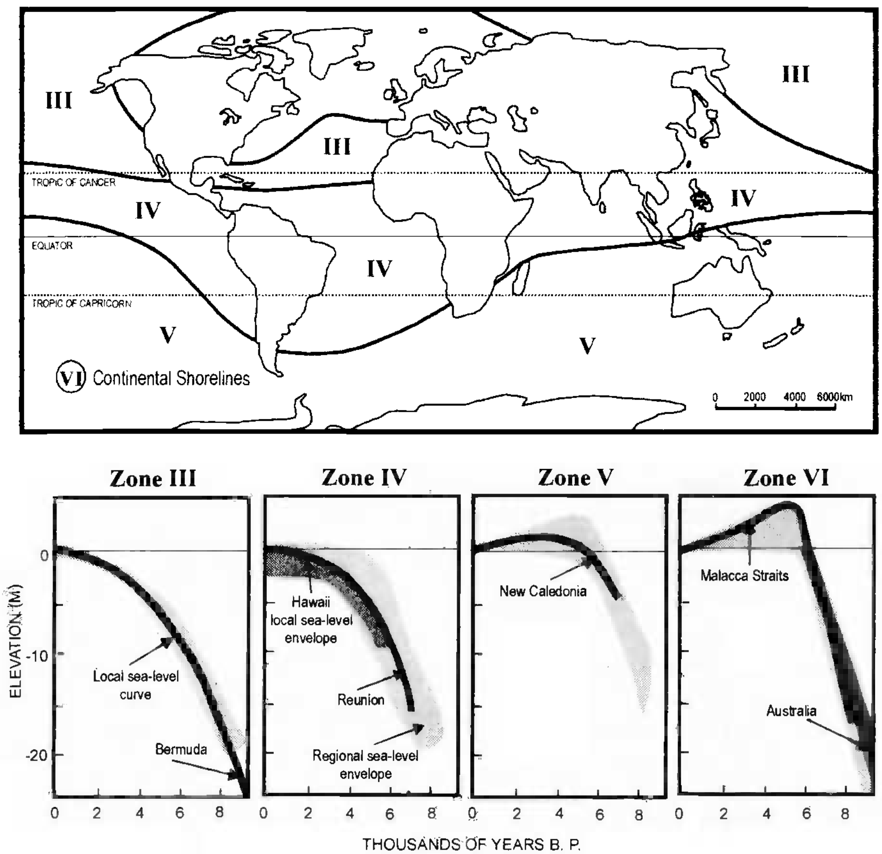


Fig. 10.3. Generalized Holocene sea level histories for equatorial regions (from Woodroffe, 1992).

Mangroves are likely to have been least disrupted during periods of major sea level change where there is a broad, coastal plain and abundant supplies of suitable sediment (Woodroffe, 1992). One such location is the southern part of the Sunda (Asian) continental shelf, centred on what is now the Malay Peninsula, where local variations in sea level are likely to have approached eustatic fluctuations because of tectonic stability. Parts of the western and southern coastline of the Malay Peninsula supported extensive mangrove forests, prior to their recent clearance for agriculture and aquaculture.

A range of evidence indicates that sea levels on the Sunda continental shelf varied by more than 100 m since the last glacial maximum (LGM, around 18,000 radiocarbon years ago) (Pirazzoli, 1991; Tjia and Fujii, 1992; Hesp *et al.*, 1998;

Hanebuth *et al.*, 2000). At times the rate of sea level rise resulting from the wastage of major ice sheets appears to have been greater – by a factor of more than five – than the maximum rates that mangrove forest can tolerate, at least according to Ellison and Stoddart (1991) and Ellison (1993). Presumably mangrove taxa on the Sunda shelf coped with high rates of sea level change in the past through lateral shifts in their distributions. There is some evidence for distributional shifts in response to past sea levels (e.g., plant remains inland of the present limits of mangrove forest), although there is an obvious danger of circularity of reasoning if the same plant fossils were also used to reconstruct sea level histories.

In those places where conditions are suitable, changes in sea level are likely to force horizontal changes in the distribution of coastal ecosystems, including mangroves. Thus, when rates of sea-level rise exceed the rate of sediment accretion and where the coastal plain is broad, relatively level and without obstructions, fringing mangrove forest is expected to respond to increased inundation by shifting inland. Similarly, during a period of marine regression and provided other factors, such as the availability of freshwater, remain constant, mangrove taxa are expected to follow the retreating intertidal zone. However, it is likely that the ability of mangrove forest to cope with varying sea levels in these ways will be severely curtailed in the future, when compared to the past. There are now few coastal areas in equatorial regions that have not been modified by humans in profound ways. Modifications in the form of agricultural and urban landscapes have the potential to act as efficient barriers to the movement inland of mangrove taxa. Furthermore, other components of global change, such as temperature rise, variations in the incidence and magnitude of rainfall, storm surges and fires and increased levels of atmospheric CO_2, will also affect the ability of mangrove taxa to deal with rising sea levels. It is possible that mangrove forest will not respond as a single, intact unit; rather mangrove taxa will respond as individuals, according to their tolerance limits, longevity and means of dispersal. Thus in the future we can expect distributional changes and new mixings of mangrove species, as a result of global change and associated impacts.

Scenarios of Anticipated Global Changes

Hulme *et al.* (1999a) maintain that humans have been the principal driver of climate change during the past 50 years. Furthermore, the IPCC (1995) conclude that increased concentrations of various greenhouse gases (carbon dioxide (CO_2), methane (CH_4), chlorofluorocarbons, nitrous oxide, ozone), emitted as a result of human activities, will lead to major global climatic change over the coming decades. At present, CO_2 is commonly thought to be the gas most responsible for greenhouse warming, originating mainly from the burning of fossil fuels, but also from forest conversion and other human practices. Atmospheric CO_2 has increased from about 280 to 353 ppm over the period 1800-1990 (IPCC, 1992; p. 68), while global mean surface air temperatures are said to have increased by 0.3-0.6°C since the late 19th century (IPCC, 1995).

Atmospheric CO_2 levels are an important forcing factor in simulations of future climate changes, based upon GCMs. GCM simulations indicate that mean global surface temperatures could rise by up to 3.5°C by 2100 as a result of increases in levels of atmospheric CO_2 (IPCC, 1995; p. 11). However, there remains great uncertainty over the precise behaviour of global climates during this period. This is because many variables, processes and feedbacks (such as the cooling affect of sulphate aerosols) remain unmodelled and because of difficulties in 'downscaling' GCM output to regional levels (Hulme *et al.*, 1999b). Thus GCMs tend to concentrate on changes in a few principal variables (such as levels of mean temperatures, precipitation and soil moisture), which may have less profound influences on ecosystems in general than, for example, changes in the frequency of extreme events. For example, no GCM is as yet able to simulate the impact of global climatic change on the El Niño Southern Oscillation (ENSO) system, so important for causing short-term perturbations in climate and ecological systems over large parts of the tropics, to a convincing level. Furthermore, GCMs currently lack the resolution to make worthwhile, long-range projections of changes in the activity of tropical cyclones (Saunders, 1999; p. 3472).

Downscaling of model simulations is particularly a problem when output and their associated impact assessments are applied to equatorial ecosystems because, being based on a fixed grid of degrees of longitude and latitude, GCMs are most coarsely resolved along the equator, where distances between degrees of longitude are greatest. Verification of GCM output is also most problematic at low latitudes owing to the relatively few meteorological and palaeoclimatic data that can be used as a basis for comparison.

In a rare example of an application of a GCM (MAGICC) to the extended tropics (30°N to 30°S), Hulme and Viner (1995) assumed a 'best' estimate for global warming of 1.76°C (which they state will be reached in the 2060s). On this basis, they predict that warming will generally be higher over tropical lands than oceans, and that land areas of the subtropics and mid-latitudes will be warmer than those nearer the equator. Their calculations suggest that temperatures over much of Southeast Asia, equatorial Africa and Central America will increase by less than 2°C, compared to over 3°C for the African subtropics and central-south Asia. They predict that annual precipitation will be reduced over much of Amazonia, southern and western Africa, and central and western Australia. However, increases in annual precipitation are predicted for a zone extending from the eastern Sahel of Africa into India.

Hulme and Viner (1995; p. 4) regard MAGICC as a 'simple climate model'. However, climate change scenarios from the model approximate to those derived for the tropics from an ensemble (mean) of four runs of HadCM2 (each run commencing with different initial conditions) and a single run of HadCM3 (Hulme *et al.*, 1999b). HadCM2 is described in full in Mavromatis and Jones (1999). The model has a horizontal resolution of 2.5° latitude x 3.5° longitude, which on the equator translates to a grid square with dimensions 278 km north-south x 417 km east-west. The ocean and atmosphere are divided into, respectively, 20 and 19 layers, with the depth of layers in the ocean becoming less as the surface is approached. HadCM3 is an improved version of HadCM2 (Hulme *et al.*, 1999b) and is fully described in Gordon *et al.* (in press). The main differences are a finer resolution of the ocean component, the inclusion of radiative forcing by aerosols and the alteration of some parameters and algorithms (see Pope *et al.*, 2000). Both models utilized the standard World Meteorological Office (WMO) normal period of averaged climate (1961-1990) as a benchmark and are forced by an annual increase in levels of atmospheric CO_2 of around 1 %. This resulted in a total increase in levels of atmospheric CO_2, relative to the period from 1961 to 1990, of around 119 % (HadCM2) and 92 % (HadCM3) by 2100 (the '2080s'). Annual temperature change output for equatorial regions from the two models is similar (summarized in Table 10.1). HadCM2 yields a slightly higher estimate of eustatic sea level rise than HadCM3 (41 cm versus 40 cm by 2100). Both models predict accelerating sea level rise over the present century, however, rising to around 0.5 cm year^{-1} during the period from 2070 to 2099. Such a rate of rise would be sufficient to cause significant stress to extensive areas of mangrove forest, at least according to the conclusions reached by Ellison and Stoddart (1991) and Ellison (1993).

Table 10.1. Climate change scenarios based on runs of the Hadley Centre's HadCM2 and HadCM3 ocean-atmosphere coupled GCMs (from Hulme *et al.*, 1999b). The benchmark is the WMO standard (i.e., data averaged for the period 1961-1990). Sea level rise is due to a combination of thermal expansion and glacier melt.

	HadCM2 (Ensemble mean)			HadCM3		
	2025 *(2020s)*	2055 *(2050s)*	2085 *(2080s)*	2025 *(2020s)*	2055 *(2050s)*	2085 *(2080s)*
Range of annual temperature changes (°C) for 30 °N to 30 °S	0 – 3	0 – 4	0 – >4	0 – 3	0 – 4	0 – >4
Eustatic sea level (m)	0.11	0.25	0.41	0.11	0.24	0.40

Uncertainties over future climates are partly expressed through different GCMs generating dissimilar climate change scenarios (e.g., Watterson *et al.*, 1999). A source of some uncertainty is the actual level of dependence of global temperature on concentrations of atmospheric CO_2 (e.g., Calder, 1999). The probability of factors other than greenhouse gases forcing climate – and hence sea level – changes in coming years is a source of further uncertainty. Indeed, climatic changes resulting from changes in land use and land cover will be of major significance, and perhaps the single largest influence on global climate change for some decades to come (Vitousek, 1994). Non-greenhouse effects on global climates have remained outside the consideration of even the most sophisticated GCMs.

Extrapolations of recent trends, output of regional and process-based models and intuition suggest that we should expect some changes to physical environments of equatorial coastal areas that are not currently modelled by GCMs. For example, on the basis of a high resolution, hurricane-prediction model, Knutson *et al.* (1998) and Knutson and Tuleya (1999) maintain that tropical storms are likely to be more vigorous as a result of global warming. Furthermore, climate models also suggest more intense but less frequent precipitation (Saunders, 1999). When combined with higher sea levels and more vigorous cyclonic activity, periods of more intense rainfall greatly increase the probability of flooding in coastal areas.

Predicted impacts on equatorial coastlines

Until recently, the state-of-the art for predictions of ecosystem response to climate change was largely limited to estimates of ecological change under some future equilibrium climate (Neilson, 1993). However, numerous studies over the past decade have tried to link equilibrium vegetation models with the output from transient GCMs (e.g., Foley *et al.*, 1998; Neilson and Drapek, 1998). Hybrid v4.1 is a complex example of a dynamic global vegetation model (DGVM), which uses as input transient output from a GCM (Friend *et al.*, 1997). When run alongside HadCM2 and HadCM3 (forced by an increase in levels of atmospheric CO_2 of around 1 % $year^{-1}$), Hybrid v4.1 predicts some replacement of tropical forests and grasslands by, respectively, grasslands and deserts (White *et al.*, 1999) over the present century. The model also predicts that these changes will be most profound from the 2050s, as a result of shifts in temperature and precipitation.

Hybrid v4.1, operating at the biome level, lacks the resolution required to make predictions for formations such as mangrove forest. However, the importance of temperature minima and frost as limits on distributions of taxa indicate that some poleward shifts of mangrove forest are likely as a result of global warming (Field, 1995). Such shifts in distributions will, of course, depend upon the availability of 'vacant', suitable ecological space into which mangrove taxa can disperse, and therefore upon local environmental conditions. However, ecological instability could cause a lessening of the breakwater-effect of mangrove forests and therefore add to the vulnerability of coastal areas to flooding, under a scenario of rising sea levels and a greater intensity of storms.

As with anticipating future changes in forcing factors, large uncertainties exist in predicting impacts of global changes, in part because of the high potential for variability at the local scale. For example, Wolanski and Chappell (1996) compared the observed and simulated behaviour of estuaries in three different geomorphological, fluvial and tidal settings in tropical, northern Australia. Different responses in sediment dynamics, river channel dimensions and vegetation to the same rise in sea level were apparent. One geomorphological feature that appears to have an important influence on response to sea level changes in the three environmental settings studied is channel form, through its effect on tidal velocities and therefore the balance between erosion and sedimentation.

Wolanski and Chappell's (1996) work demonstrates the complex influence of physiographic conditions on response to sea level change, as their study sites were virtually uninhabited by humans. However, humans are probably the greatest imponderables in the prediction of responses to environmental changes in coastal areas. In a relatively early attempt to assess the socio-economic and ecological implications of eustatic sea level rise, Hoozemans *et al.* (1993) and Hoozemans and Hulsbergen (1995) conducted a Global Vulnerability Assessment (GVA), based upon the conclusions reached by the IPCC's Coastal Zone Management Sub-group (IPCC CZMS, 1990, 1992). The CZMS estimated that eustatic sea levels would rise by around 1 m over the period from 1990 to 2100 (their prediction is now regarded as an overestimate). On this basis, and assuming no other changes, Hoozemans *et al.* (1993) predicted that the number of people affected by flooding would increase by about one third, with the coastlines of South Asia and Africa together with oceanic islands among the locations most effected. Baarse (1995) extended the GVA to include the increasing risk of flood to the population within the existing coastal plain and estimated that the number of people directly affected by storm surges would triple in a typical year, given a sea level rise of 1 m.

Most recently, Nicholls *et al.* (1999) carried out an assessment that was more finely resolved, spatially and temporally, and was based upon a more complete data set and a lower estimated rise in eustatic sea levels than earlier studies. Their assessment combined the HadCM2 and HadCM3 scenarios described previously with data on population increase and changes in gross national product (as a guide to the ability to pay for protection schemes) in a coastal flood model. The model incorporates processes such as sediment accretion and ecosystem migration (Nicholls *et al.*, 1999; p. 576), but assumes a uniform coastal slope and does not include any site-specific information, such as local hydrological conditions and the availability and type of sediments. Nor does the model accommodate the differences in resilience of people living in affected areas, many of whom are expected to have developed ways of coping with frequent and at times severe floods. Output from the model should therefore not be taken too literally, although simulations do provide an indication of the level of regional differences in vulnerability.

Output from the model indicates that the average number of people affected by floods each year (10 million in 1990) will rise to 36 million in the 2080s, assuming maintenance of flood protection at 1990 levels and no eustatic sea level changes (although the model allows for some subsidence). With evolving protection under conditions of stable eustatic sea levels, the number first rises to 27 million in the 2050s, before falling to 13 million by the 2080s. These numbers are increased greatly with HadCM2 and HadCM3 predicted rises in sea level of around 40 cm by 2100: 237 million people are anticipated to be affected if protection is maintained at 1990 levels; 93 million if protection evolves as sea levels rise. At a regional scale, the coastal areas most vulnerable to hazards are predicted to be equatorial Africa and southern and Southeast Asia (Fig. 10.4). Together with the southern Mediterranean, the model predicts that these areas will account for more than 90 % of people affected by flooding in the 2080s, irrespective of the protection scheme (Nicholls *et al.*, 1999; p. 580). South and Southeast Asia are said to be particularly vulnerable, because of high population levels on low-lying deltas (Fig. 10.5) and coastal plains, with well over 50% of all people affected globally living in these two regions. The percentage is even greater under a scenario of evolving protection, because the relatively low GNPs in

South and Southeast Asia, and therefore a general shortage of funds to pay for coastal protection, will constrain the range of responses that are possible. An inability to pay for protection schemes that involve some armouring of the coastline, and therefore that constrain lateral shifts of intertidal ecosystems during a period of rising sea level, may be of some benefit to coastal wetlands in equatorial regions. Moreover, Nicholls *et al.* (1999) predict that coastal populations in developed countries may be least vulnerable to a sustained rise in sea level, whereas actual losses of coastal wetland habitat is likely to be greatest.

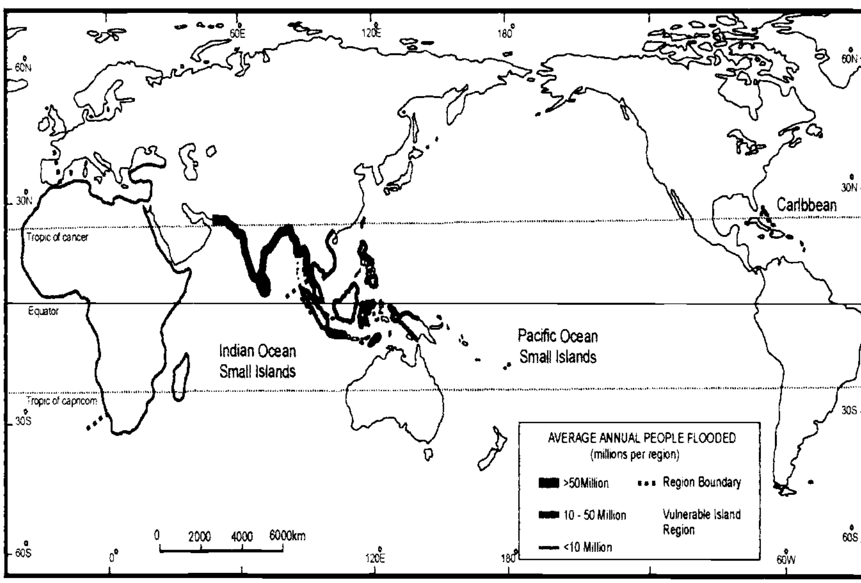


Fig. 10.4. Coastal areas most vulnerable to global changes (based on Nicholls *et al.*, 1999).

Nicholls *et al.* (1995) provide a synthesis of the implications of global climate changes for coastal areas in south and Southeast Asia. The synthesis relied upon climate and sea level change scenarios outlined in the IPCC's First Scientific Assessment (IPCC, 1992). The authors conclude that coastal areas in the region will change in coming decades, with or without climate change, but suggest that climate change is probable and likely to exacerbate current problems of flooding, salinization and losses of wetlands and coastal forests. They recommend an integrated, international response that aims to address both current and anticipated problems.

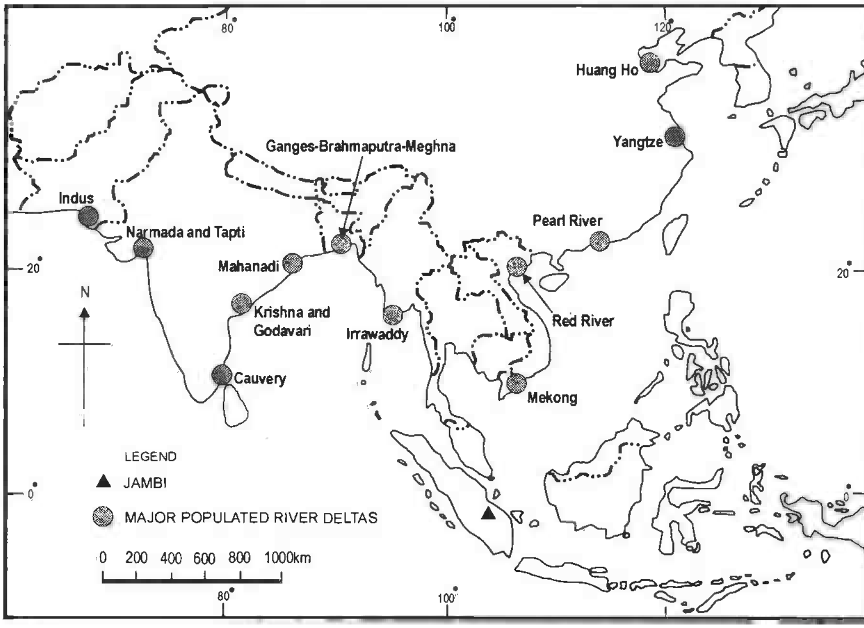


Fig. 10.5. Locations of densely populated coastal deltas in south and Southeast Asia (from Nicholls *et al.*, 1995) and of Jambi on the island of Sumatra.

Global change and forested coastlines in Sumatra

Nicholls *et al.* (1999) emphasize the uncertainty that is inherent in their GVA. Some indication of the actual complexity that exists in the local response to changing coastal environments is provided by the following example from eastern Sumatra, based in part on a number of visits by the authors to coastal areas of Jambi.

Wetlands are extensive over the coastal plain and along the lower reaches of river valleys in eastern Sumatra and, on the basis of very few radiocarbon dates, would appear to date to the period after the mid-Holocene regional sea level maximum. The majority of coastal wetlands supported forests until relatively recently, with major deforestation dating in some parts to the 19th century, but mainly to the last century and particularly since the late-1980s (Furukawa, 1994). Laumonier (1997), on the basis of land-use maps and satellite data, estimated that ‘practically intact’ mangrove forest covers about 200,000 ha of these wetlands, with depleted or secondary forms accounting for a further 100,000 to 150,000 ha. Towards the upper limit of tidal influence, mangroves in Sumatra often grade into *Nypa fruticans*-dominated, brackish water forest. With further reductions in salinity, freshwater peatswamp forest predominates, which in many places has formed over palaeosols (red earths). Both brackish water and peatswamp forests have been extensively modified by humans and converted to agricultural land. The extent of brackish water forests is difficult to ascertain, because of its fragmented occurrence, although Laumonier (1997) estimates that peatswamp forest covers around 38,000 km^2, or something like 8 % of the total area of Sumatra.

Peat and alluvium on the coastal plain of Jambi province are believed to date to episodic progradation that has taken place during the past 6000 years or so. The last major episode followed the termination of a minor marine transgression, around 1500 years ago, during which the present drainage system and areas of deltaic deposits were formed (Furukawa, 1994; p. 29). Just how much of the present coastal plain dates to this period is unclear, because the relatively few sedimentary studies that have been conducted lack satisfactory chronological control. It could be a substantial area, however, based on the location of a major temple complex, 'Muara Jambi', some 70 km from the present mouth of the Sungai (River) Batanghari. Excavations in the mid 1970s revealed a style of deposit chambers and artefacts that date parts of the site to the late 9th and early 10th centuries (Miksic, 1989; p. xii). Translated, 'Muara Jambi' means 'mouth of the Jambi'. If referring to the proximity of the mouth of the main river (Batanghari) around the time the temples were being constructed, as seems likely, the use of the name perhaps attests to considerable progradation over the past 1000 years or so and a rapid rate of horizontal retreat of the intertidal zone. Estimates place the modern rate at which the coastal plain of eastern Sumatra is being extended seaward to be of the order of 30 to 125 m $year^{-1}$ (Macnae, 1968; Sobur *et al.*, 1977) and more than 200 m $year^{-1}$ in some deltaic areas (Guelorget *et al.*, 1996). At least some of the sediments that are accumulating in coastal areas are likely to be terrigenous and to have been eroded from interior areas as a result of major changes in land use. Much of Sumatra is presently undergoing an 'agricultural revolution', in which areas of forest and existing agricultural land are being transformed, mainly to oil palm and *Acacia* plantations.

According to documentary evidence, the majority of the coastal plain in Jambi was virtually unoccupied as recently as the 1950s. The situation changed during the late 1970s, when transmigration and plantation schemes were first introduced. The extent of the schemes was increased and there was some diversification of crops, mainly as a result of reduced rice productivity, during the 1980s. The local landscape has been massively transformed since 1989 (Furukawa, 1994; p. 48), however, as a result of the clearance of a large amount of the remaining coastal and peatswamp forests and its replacement with plantation agriculture.

Lee and Couturier (unpublished data) have quantified many of the changes during the period from 1989 to 1999, based on a comparison of Landsat and mosaiced SPOT satellite data. The data, some of which are presented in Table 10.2, are for a 14,000 km^2 swathe of land that includes a large part of the coastal plain of Jambi province. Aside from an elimination of intact peatswamp forest and an expansion of plantation agriculture (40 % increase), the data indicate that the area of mangrove forest (the category probably includes some brackish water forest) declined by more than 25 % during the 10 year study period. Thus, over a period for which there is rapid progradation of the coastal zone, possibly partly as a result of the mobilization of soil by land use changes inland, the absolute extent of mangrove forest has declined. Two processes stand out as offering an explanation for this. First, progradation and mangrove colonization are either patchy, and restricted to areas close to the mouths of estuaries, where supplies of freshwater and sediment are greatest, or proceeding out of tandem. There is some anecdotal evidence of the latter, or mangrove colonization lagging sediment accretion. An interview with the headman of a village at the mouth of one of the rivers draining the area maintained that increased siltation had contributed to reduced catches of fish and accelerated seaward extensions of mangrove thicket (Tan, 1999; p. 67). The figure for the rate of extension he gave was approximately 4 to 5 m year^{-1}, which is far less than the general estimates of the rate of progradation cited earlier. Second, mangrove is being cleared on its landward margins for agriculture and timber, which serves a range of needs in the local economy, including the construction of buildings and walkways (villages in the area are still predominantly wood-built) elevated well above the level of high tides, boats for fishing and transportation and permanent and temporary fish traps.

Table 10.2. Land use change in eastern Sumatra, 1989 to 1999, based on a comparison of Landsat (1989) and mosaiced SPOT (1999) satellite data. See Fig.10.5 for location of study. From Lee and Couturier (unpublished data).

Land use	1989 area (ha)	1999 area (ha)	Direction of change (%)
Intact peatswamp forest	18,804	0	Decline (100)
Logged peatswamp forest	260,132	167,666	Decline (35.5)
Burned peatswamp forest	0	60,270	Increase
Mangrove	7177	5166	Decline (28)
Plantation agriculture	212,579	299,639	Increase (40.1)

Although the coastline of eastern Sumatra is generally classed as a low energy environment, storm surges do occur and these may have serious consequences, especially when they coincide with a high tide. The diurnal tidal range is generally around 3 to 4 m (Furukawa, 1994), which results in the zone of saline intrusion being pushed far inland along the low-lying drainage network, even under non-storm conditions. Continued clearance of mangrove forest and seaward extensions of agricultural land will jeopardize the chances of remaining mangrove taxa of surviving global changes in the future, reduce the ability of mangrove ecosystems to protect against storm surges, sediment scouring and saltwater intrusion and thus over large areas threaten the livelihoods of farmers and fishermen.

Conclusion

Even under stable climate conditions, living in and obtaining a livelihood from coastal regions is likely to become more hazardous. This is because of the expected increases in levels of human population and a greater susceptibility to flooding and salinization, as a result of compaction and subsidence (Woodroffe, 1995), surface sealing and loss of buffer zones between the coast and settled areas. These 'other' variables are most apparent in the developing (largely equatorial) world, where rates of population increase are highest and protection schemes and mitigation strategies among the most poorly advanced. A scenario involving any form of environmental stability is not realistic, of course. We live in a highly dynamic world, both economically and environmentally. This dynamism is as much a characteristic of coastal regions in the tropics as it is of higher latitudes. The nature of this dynamism and the myriad of associated feedbacks, non-linearities, impacts and responses may be difficult to predict at present to a high level of certainty. However, it is certain that future changes to coastal environments, whatever their cause, initial starting point and subsequent trajectory of evolution, will threaten livelihoods and therefore increase vulnerability to hazards.

Levels of hazard will be least in economically developed countries, because of a combination of relatively low rates of population increase and an ability and willingness to pay for protection schemes. Where these schemes involve armouring of the coastline they will threaten existing ecosystems, because lateral adjustments to sea level rise will be constrained and because the power of the sea may merely be deflected to, and therefore enhance erosion of, unprotected parts of the shoreline. But there are relatively few economically developed countries in equatorial regions and here the situation is likely to be very different. Many developing countries are unable to meet the costs of protecting against the threat posed at present by coastal processes such as storm surges, in part because of the sheer extent of populated coastlines. In these cases, a precautionary approach that involves major investments in schemes that aim to provide some protection against environmental changes that may occur in the future is unlikely, without major outside assistance. This may mean that there is a greater potential for formations such as mangrove forest to adjust laterally to sea level changes in the future, when compared to those of more temperate latitudes. However, this is only likely in those locations where there is little or no intensive human activity on the coastal plain. Much more likely is that, in many places, existing coastal forests will be squeezed between rising sea levels and agricultural land, while farmers will have to contend with higher instances of damaging floods and increased salinity. The absence of adaption strategies, such as the adoption of salt-tolerant crops and associated technologies, will force the abandonment of formerly productive land. Displacement of people will increase population pressures inland. The numbers affected could be considerable; Nicholls *et al.* (1995) estimate that nearly 15 million people in Bangladesh and at least 2 million people in Indonesia could be displaced by a 1 m rise in sea level.

Nicholls *et al.* (1999) propose three precautionary strategies to reduce vulnerability to sea level change-induced hazards in coastal areas. Of these, planned retreat (in which the most vulnerable areas are zoned as unsuitable for development) is perhaps the most feasible in equatorial regions, but only for the least densely populated coastlines. A second strategy - proofing accommodation against flooding - is already practised in some parts of the developing world, for example by constructing buildings and walkways on piles, well above water levels. However, the practice can be very demanding of timber and other natural resources and, as seen in Sumatra, may itself contribute to a decline in extent of mangrove forest, and hence to the degradation of a natural form of coastal protection.

The prognosis for forested coastlines in equatorial regions is poor and is unlikely to improve while coastal environments remain locations for the juxtaposition of dynamic natural environments with relatively static, ecologically disruptive, economic activities. Formations such as mangrove forest face the combined threat of global change-induced variations in environmental conditions and clearance and over-exploitation by humans. By degrading mangrove forests, however, humans will remove one of the few effective means of protecting coastal areas against the impacts of global change. Where removal of coastal forests proceeds along with the construction of new drainage channels and the canalization of existing rivers, impacts of flood surges and intrusions of saline water are likely to be felt sooner and over larger areas than on coastal plains with low human impact. By their actions, humans may therefore enhance the possibility of catastrophic events taking place in the future, and therefore their own vulnerability to cumulative hazards.

One possible strategy that humans might adopt to mitigate the impacts of global change on equatorial coastlines, and one that – surprisingly – is not considered by Nicholls *et al.* (1999), is a co-ordinated effort to maintain, restore and promote at least a fringing band of coastal forest, especially along the most vulnerable coastlines. If that forest were mangrove, then it would provide combined coastal protection, a source of natural resources and a habitat for many species of wildlife (Kaly and Jones, 1998). The strategy is among those recommended by Tri *et al.* (1998) who, on the basis of Vietnam's densely populated coastal zone, maintains that mangrove forests are a cost effective means of protecting coastlines, including those that are presently fronted by sea dykes.

Acknowledgements

The authors would like to thank Roy C. Sidle and an anonymous referee for helpful comments on an earlier draft of this chapter. Information on coastal areas and populations in Jambi, which the authors refer to, was gathered as part of a LOICZ (Land-Ocean Interactions in the Coastal Zone) Research Project under the International Geosphere-Biosphere Programme (IGBP). Full funding for this project was provided by National University of Singapore (NUS) – research grant #RP 3981034.

References

Anderson, J.A.R. (1964) The structure and development of peat swamps of Sarawak and Brunei. *The Journal of Tropical Geography* 18, 7-16.

Anderson, J.A.R. (1983) The tropical peat swamps of Western Malesia. In: Gore, A.J.P (ed.), *Ecosystems of the World 4B*. Elsevier Scientific Publishing Company, New York, pp. 181-199.

Anderson, J.A.R. and Muller, J. (1975) Palynological study of a Holocene peat and a Miocene coal deposit from NW Borneo. *Review of Palaeobotany and Palynology* 19, 291-351.

Baarse, G. (1995) Development of an Operational Tool for Global Vulnerability Assessment (GVA): Update of the Number of People at Risk due to Sea-Level Rise and Increased Flood Probabilities. *CZM-Centre Publications No 3*. Ministry of Transport, Public Works and Water Management, The Hague, the Netherlands.

Bijlsma, L., Ehler, C.N., Klein, R.J.T., Kulshrestha, S.M., McLean, R.F., Mimura, N., Nicholls, R.J., Nurse, L.A., Perez Nieto, H., Stakhiv, E.Z., Turner, R.K. and Warrick, R.A. (1996) Coastal zones and small islands. In: Watson, R.T., Zinyowera, M.C. and Moss, R.H. (eds), *Impacts, Adaptations and Mitigation of Climate Change: Scientific-Technical Analyses*. Cambridge University Press, Cambridge, pp. 289-324.

Blasco, F., Saenger, P. and Janodet, E. (1996) Mangroves as indicators of coastal change. *Catena* 27, 167-178.

Burton, I., Kates, R.W. and White, G.F. (1993) *The Environment as Hazard*, 2nd edn. The Guilford Press, New York.

Calder, N. (1999) The carbon dioxide thermometer and the cause of global warming. *Energy and Environment* 10, 1-18.

Chapman, V.J. (1976) *Mangrove Vegetation*. J.Cramer, Vaduz.

Clough, B.F. (1992) Primary productivity and growth of mangrove forests. In: Robertson, A.I. and Alongi, D.M. (eds), *Tropical Mangrove Ecosystems*. American Geophysical Union, Washington, DC, pp. 225-249.

Duke, N.C. (1992) Mangrove floristics and biogeography. In: Robertson, A.I. and Alongi, D.M. (eds), *Tropical Mangrove Ecosystems*. American Geophysical Union, Washington DC, pp. 63-100.

Ellison, J.C. (1993) Mangrove retreat with rising sea-level. *Bermudan Estuarine and Coastal Marine Science* 37, 75-87.

Ellison, J.C. and Stoddart, D.R. (1991) Mangrove ecosystem collapse during predicted sea-level rise: Holocene analogues and implications. *Journal of Coastal Research* 7, 151-165.

Field, C.D. (1995) Impact of expected climate change on mangroves. *Hydrobiologia* 295, 75-81.

Foley, J.A., Levis, S., Prentice, I.C., Pollard, D. and Thompson, S.L. (1998) Coupling dynamic models of climate and vegetation. *Global Change Biology* 4, 561-579.

Friend, A.D., Stevens, A.K., Knox, R.G. and Cannell, M.G.R. (1997) A process-based terrestrial biosphere model of ecosystem dynamics (Hybrid v.3.0). *Ecological Modelling* 95, 249-287.

Furukawa, H. (1994) *Coastal Wetlands of Indonesia: Environment, Subsistence and Exploitation*. Kyoto University Press, Japan.

Furukawa, H. and Sabiham, S. (1985) Agricultural landscapes in the wetlands of the Batang Hari River basin. Part 1. Landforms and sedimentary stratigraphy. *Southeast Asian Studies* 23, 3-37.

Gares, P.A., Sherman, D.J. and Nordstrom, K.F. (1994) Geomorphology and natural hazards. *Geomorphology* 10, 1-18.

Gommes, R., du Guerny, J., Nachtergaele, F. and Brinkman, R. (1997) *Potential impacts of sea-level rise on populations and agriculture*. FAO, Rome.

Gordon, C., Cooper, C., Senior, C., Banks, H., Gregory, J.M., Johns, T.C., Mitchell, J.F.B. and Wood, R. (2000) The simulation of SST, sea ice extents and ocean heat transports in a version of the Hadley Centre coupled model without flux adjustments. *Climate Dynamics* 16, 147-168.

Guelorget, O., Martin, J.L., Lefebvre, A., Courties, C., Perthuisot, J.P., Supangat, A., Fuchs, J. and Suastika, M. (1996) Open sea paralic ecosystems south of Java Sea: environmental approach by flow cytometry. *Journal of Coastal Research* 12, 256-270.

Hanebuth, T., Stattegger, K. and Grootes, P.M. (2000) Rapid flooding of the Sunda Shelf: A late-glacial sea-level record. *Science* 288, 1033-1035.

Hesp, P.A., Chang, C.H., Hilton, M., Chou, L.M. and Turner, I. (1998) A first tentative Holocene sea-level curve for Singapore. *Journal of Coastal Research* 14, 308-314.

Hoozemans, F.M.J. and Hulsbergen, C.H. (1995) Sea-level rise: a world-wide assessment of risk and protection costs. In: Eisma, D. (ed.), *Climate Change Impact on Coastal Habitation*. Lewis Publishers, London, pp. 137-163.

Hoozemans, F.M.J., Marchand, M. and Pennekamp, H.A. (1993) *A Global Vulnerability Analysis: Vulnerability Assessment for Population, Coastal Wetlands and Rice Production on a Global Scale*, 2nd edn. Delft Hydraulics, The Netherlands.

Hulme, M. and Viner, D. (1995) A climate change scenario for assessing the impact of climate change on tropical rain forests. *Report prepared for the Climatic Research Unit for WWF (US)*. Climatic Research Unit, University of East Anglia, Norwich, UK.

Hulme, M., Barrow, E.M., Arnell, N.W., Harrrison, P.A., Johns, T.C. and Downing, T.E. (1999a) Relative Impacts of Human Induced Climate Change and Natural Variability. *Nature* 397, 688-691.

Hulme, M., Mitchell, J., Ingram, W., Lowe, J., Johns, T., New, M. and Viner, D. (1999b) Climate change scenarios for global impacts studies. *Global Environmental Change – Human and Policy Dimensions* 9, S3-S19.

IPCC (1992) *Climate Change: The IPCC 1990 and 1992 Assessments*. WMO/UNEP, Cambridge University Press, Cambridge, UK.

IPCC (1995) *Climate Change 1995: The Science of Climate Change Summary for Policymakers*. WMO/UNEP, Cambridge University Press, Cambridge, UK.

IPCC (1997) *The Regional Impacts of Climate Change: An Assessment of Vulnerability. Summary for Policymakers*. A Special Report of IPCC Working Group II WMO/UNEP, Cambridge University Press, Cambridge, UK.

IPCC CZMS (1990) Strategies for adaptation to sea level rise. *Report of the Coastal Zone Management Subgroup*. IPCC Response Strategies Working Group, Rijkswaterstaat, The Hague, The Netherlands.

IPCC CZMS (1992) Global climate change and the rising challenge of the sea. *Report of the Coastal Zone Management Subgroup*. IPCC Response Strategies Working Group, Rijkswaterstaat, The Hague, The Netherlands.

Kaly, U.L. and Jones, G.P. (1998) Mangrove restoration: a potential tool for coastal management in tropical developing countries. *Ambio* 27, 656-661.

Klein, R.J.T. and Nicholls, R.J. (1999) Assessment of coastal vulnerability to sea-level rise. *Ambio* 28, 182-187.

Knox, G.A. (1986) *Estuarine Ecosystems: A Systems Approach*, Volume 1. CRC Press, Boca Raton, Florida.

Knutson, T.R. and Tuleya, R.E. (1999) Increased hurricane intensities with CO_2-induced warming as simulated using the DFDL hurricane prediction system. *Climate Dynamics* 15, 503-519.

Knutson, T.R., Tuleya, R.E. and Kurihara, Y. (1998) Simulated increase of hurricane intensities in a CO_2-warmed climate. *Science* 279, 1018-1020.

Laumonier, Y. (1997) *The Vegetation and Physiography of Sumatra*. Kluwer Academic Publishers, Dordrecht.

Lugo, A.E. and Patterson-Zucca, C. (1977) The impact of low temperature stress on mangrove structure and growth. *Tropical Ecology* 18, 149-161.

Macnae, W. (1968) A general account of the flora and fauna of mangrove swamps and forests in the Indo-West-Pacific region. *Advances in Marine Biology* 6, 73-270.

Massel, S.R., Furukawa, K. and Brinkman, R.M. (1999) Surface wave propagation in mangrove forests. *Fluid Dynamics Research* 24, 219-249.

Mavromatis, T. and Jones, P.D. (1999) Evaluation of HadCM2 and direct use of daily GCM data in impact assessment studies. *Climate Change* 41, 583-614.

Miksic, J.N. (1989) Introduction. In: Schnitger, F.M. (ed.), *Forgotten Kingdoms in Sumatra*. Oxford University Press, Oxford, UK.

Neilson, R.P. (1993) Transient ecotone response to climate change – some conceptual and modelling approaches. *Ecological Applications* 3, 385-395.

Neilson, R.P. and Drapek, R.J. (1998) Potentially complex biosphere responses to transient global warming. *Global Change Biology* 4, 505-521.

Nicholls, R.J. (1995) Coastal megacities and climate change. *GeoJournal* 37, 369-379.

Nicholls, R.J., Mimura, N. and Topping, J.C. (1995) Climate change in South and Southeast Asia: some implications for coastal areas. *Journal of Global Environment Engineering* 1, 137-154.

Nicholls, R.J., Hoozemans, F.M.J. and Marchand, M. (1999) Increasing flood risk and wetland losses due to global sea-level rise: regional and global analyses. *Global Environmental Change – Human and Policy Dimensions* 9, S69-87.

Phillips, S., Rouse, G.E. and Bustin, R.M. (1997) Vegetation zones and diagnostic pollen profiles of a coastal peat swamp, Bocas del Toro, Panamá. *Palaeogeography, Palaeoclimatology, Palaeoecology* 128, 301-338.

Pirazzoli, P.A. (1991) *World Atlas of Holocene Sea Level Changes*. Elsevier Oceanography Series 58. Elsevier, Amsterdam.

Pope, V.D., Gallani, M.L., Rowntree, P.R. and Stratton, R.A. (2000) The impact of new physical parameterizations in the Hadley Centre climate model – HadCM3. *Climate Dynamics* 16, 123-146.

Saunders, M.A. (1999) Earth's future climate. *Philosophical Transactions of the Royal Society of London Series A* 357, 3459-3480.

Scheidegger, A.E. (1994) Hazards: singularities in geomorphic systems. *Geomorphology* 10, 19-25.

Short, F.T. and Neckles, H.A. (1999) The effects of global change on seagrass. *Aquatic Botany* 63, 169-196.

Smith, T.J. III (1992) Forest structure. In: Robertson, A.I. and Alongi, D.M. (eds), *Tropical Mangrove Ecosystems*. American Geophysical Union, Washington, DC, pp. 101-136.

Smith, T.J. III and Duke, N.C. (1987) Physical determinants of inter estuary variation in mangrove species richness around the tropical coastline of Australia. *Journal of Biogeography* 14, 9-19.

Snedaker, S.C. (1995) Mangrove and climate change in the Florida and Caribbean regions: scenarios and hypotheses. *Hydrobiologia* 295(1-3), 43-49.

Sobur, A.S., Chambers, M.J., Chambers, R., Damapoli, S.H. and Hanson, A.J. (1977) *Remote sensing applications in the Southeast Sumatera coastal environment*. IPB, Bogor, Indonesia.

Tan, P.M. (1999) The political ecology of coastal resource access in a Sumatran fishing village. Unpublished Honours Thesis, National University of Singapore, Singapore.

Thom, B.G. (1982) Mangrove ecology: a geomorphological perspective. In: Clough, B.F. (ed.), *Mangrove ecosystems in Australia, structure, function and management.* Australia National University, Canberra, pp. 3-17.

Tjia, H.D. and Fujii, S. (1992) Late Quaternary shorelines in Peninsular Malaysia. In: Syed Abdullah, S.M. (ed.), *The Coastal Zone of Peninsular Malaysia.* Pernerbit Universiti, Malaysia, pp. 28-41.

Tri, N.H., Adger, W.N. and Kelly, P.M. (1998) Natural resource management in mitigating climate impacts: the example of mangrove restoration in Vietnam. *Global Environmental Change – Human and Policy Dimensions* 8, 49-61.

Vitousek, P.M. (1994) Beyond global warming: ecology and global change. *Ecology* 75, 1861-1876.

Walsh, G.E. (1974) Mangroves: a review. In: Reimold, R.J. and Queen, W.H. (eds), *Ecology of Halophytes.* Academic Press, New York, pp. 51-174.

Watterson, I.G., Dix, M.R. and Colman, R.A. (1999) A comparison of present and doubled CO_2 climates and feedbacks simulated by three general circulation models. *Journal of Geophysical Research – Atmospheres* 104, 1943-1956.

White, A., Cannell, M.G.R. and Friend, A.D. (1999) Climate change impacts on ecosystems and the terrestrial carbon sink: a new assessment. *Global Environmental Change – Human and Policy Dimensions* 9, S21-S30.

Whitmore, T.C. (1984) *Tropical Rain Forests of the Far East.* Clarendon Press, Oxford, UK.

Whitmore, T.C. (1998) *An Introduction to Tropical Rain Forests*, 2nd edn. Oxford University Press, Oxford, UK.

Wigley, T.M.L. (1995) Global-mean temperature and sea level consequences of greenhouse gas stabilization. *Geophysical Research Letters* 22, 45-48.

Williams, M., Dunkerley, D., De Deckker, P., Kershaw, P. and Chappell, J. (1998) *Quaternary Environments*, 2nd edn. Arnold, London.

Wolanski, E. and Chappell, J. (1996) The response of tropical Australian estuaries to a sea level rise. *Journal of Marine Systems* 7, 267-279.

Woodroffe, C.D. (1988) Changing mangrove and wetland habitats over the past 8000 years, northern Australia and Southeast Asia. In: Wade-Marshall, D. and Loveday, P. (eds), *Northern Australia: Progress and Prospects Volume 2: Floodplains research.* North Australia Research Unit, ANU, Canberra, pp. 1-33.

Woodroffe, C.D. (1990) The impact of sea-level rise on mangrove shorelines. *Progress in Physical Geography* 14, 483-520

Woodroffe, C.D. (1992) Mangrove sediments and geomorphology. In: Robertson, A.I. and Alongi, D.M. (eds), *Tropical Mangrove Ecosystems.* American Geophysical Union, Washington DC, pp. 7-41.

Woodroffe, C.D. (1995) Response of tide-dominated mangrove shorelines in northern Australia to anticipated sea-level rise. *Earth Surface Processes and Landforms* 20, 65-85.

Woodroffe, C.D. and Grime, D. (1999) Storm impact and evolution of a mangrove-fringed chenier plain, Shoal Bay, Darwin, Australia. *Marine Geology* 159, 303-321.

Woodroffe, C.D. and Grindrod, J. (1991) Modern biogeography: the role of Quaternary environmental and sea-level change. *Journal of Biogeography* 18, 479-492.

Wright, L.D., Coleman, J.M. and Erikson, M.W. (1974) Analysis of major river systems and their deltas: morphologic and process comparisons. Global Studies Institute, Louisiana State University, *Technical Report* 156.

Chapter 11

Future Directions for Geomorphologic Hazard Analysis in Forests

Roy C. Sidle
Department of Geography, The National University of Singapore, Singapore

Areas of Increasing Focus in a Changing Environment

As climate is one of the primary driving or triggering mechanisms of many geomorphic hazards, any changes in long-term climate may alter the methods currently used to quantify hazard occurrence. For example, traditional statistical approaches that assume stationarity of climatic time series data are not appropriate for analysing future trends in rainfall and temperature properties that trigger geomorphic hazards (e.g., Kiely, 1999). Other types of trend analysis need to be employed to analyse long-term climate data. However, resource managers and planners require estimates of climate trends in areas where long-term basic data do not exist or are of poor quality. This is particularly true in developing regions of the tropics. Given the current and near future unlikelihood that global circulation models (GCMs) will produce accurate climate change predictions at the spatial scale needed for most geomorphic hazard analyses, an alternative approach is to represent climate change attributes as conditional probabilities in geomorphic models or calculations. One method is to include potential climatic trigger mechanisms in models as stochastic variables (e.g., Sidle and Wu, 1999). Best estimates of climate change scenarios can be gleaned from GCMs or other trend analyses; these can then be incorporated into stochastic models with appropriate estimates of parameter variability. Such an approach if incorporated into regional scale hazard analysis can be quite useful for resource managers. Analysis of geomorphic hazards at more detailed scales in changing climatic regimes remains problematic. For such situations, best estimates of likely patterns of the triggering climate variables will need to be combined with local knowledge of trajectories of land use and demographics to develop contemporary geomorphic hazard assessments. As noted in a number of the chapters of this volume, anthropogenic change may overshadow the potential impacts of climate change with respect to land degradation and related erosion hazards, particularly in developing or rapidly urbanizing regions.

One of the challenges related to changing forest land uses is assessing cumulative effects of these practices in time and space dimensions. Land uses in predominantly forested and rangeland areas that generate cumulative effects include timber harvesting (with subsequent forest regeneration), grazing, vegetation conversion (primarily to agriculture), recreation, roads and trails, mining, waste disposal, fire, and urban encroachment. Long-term climate change and regional patterns of atmospheric deposition represent a broader 'layer' of anthropogenic factors that may enhance the impacts of these land uses. However, a comprehensive definition of cumulative effects must recognize the interaction of natural ecosystem processes (e.g., geomorphic hazards) with the effects of anthropogenic changes distributed in time and space (Sidle and Hornbeck, 1991). Many of the current approaches to assessing cumulative effects rely on indicators of land disturbance such as 'equivalent clearcut area' and 'equivalent roaded area' (e.g., McGurk and Fong, 1995). However, such 'lumped' approaches do not address the spatial and temporal aspects of land disturbances related to geomorphic hazards. To develop better tools for forest management, spatially distributed land uses need to be incorporated into analyses and models (e.g., Bevers *et al.*, 1996).

Anthropogenic factors can alter the thresholds for initiation or acceleration of certain hazards (Fig. 11.1). Such altered thresholds may be manifested at various spatial scales particularly for hazards occurring in steep terrain where effects may cascade down through terrestrial/aquatic ecosystems. While cumulative effects may be somewhat modulated as materials move or are transformed through catchments of increasing size (Dunne, 1998; Thornton *et al.*, 2000), certain types of hazards may create new thresholds within the system (spatially removed from the original hazard) that present future risks. These considerations are especially important in developing methods to analyse geomorphic hazards that accrue as the result of distributed spatial processes. For example, processes that control stormflow generation change with scale and catchment attributes, and flow routing is affected by sedimentation and channel conditions. Additionally, landslide hazards must be assessed not only on-site, but also related to the transport of materials, often by debris flows (Dunne, 1998), as well as impacts on the flood conveyance capacity of channels. Sea level rise may inundate coastal regions, but larger scale effects of such rises include salinization and accelerated coastal and fluvial erosion. These examples illustrate that one type of hazard may exacerbate another hazard.

Emerging Methodologies and Techniques

Geomorphic hazard analysis has and will benefit from advances in remotely sensed data acquisition and generation of digital elevation models (DEMs). As remote sensing techniques improve, changes in important environmental parameters that affect natural hazards such as vegetation cover (Hugh and Lambin, 2000), rainfall intensity (Marzano *et al.*, 2001), sediment distribution (Pickup and Marks, 2001), soil moisture (Moran *et al.*, 2000), and snow water

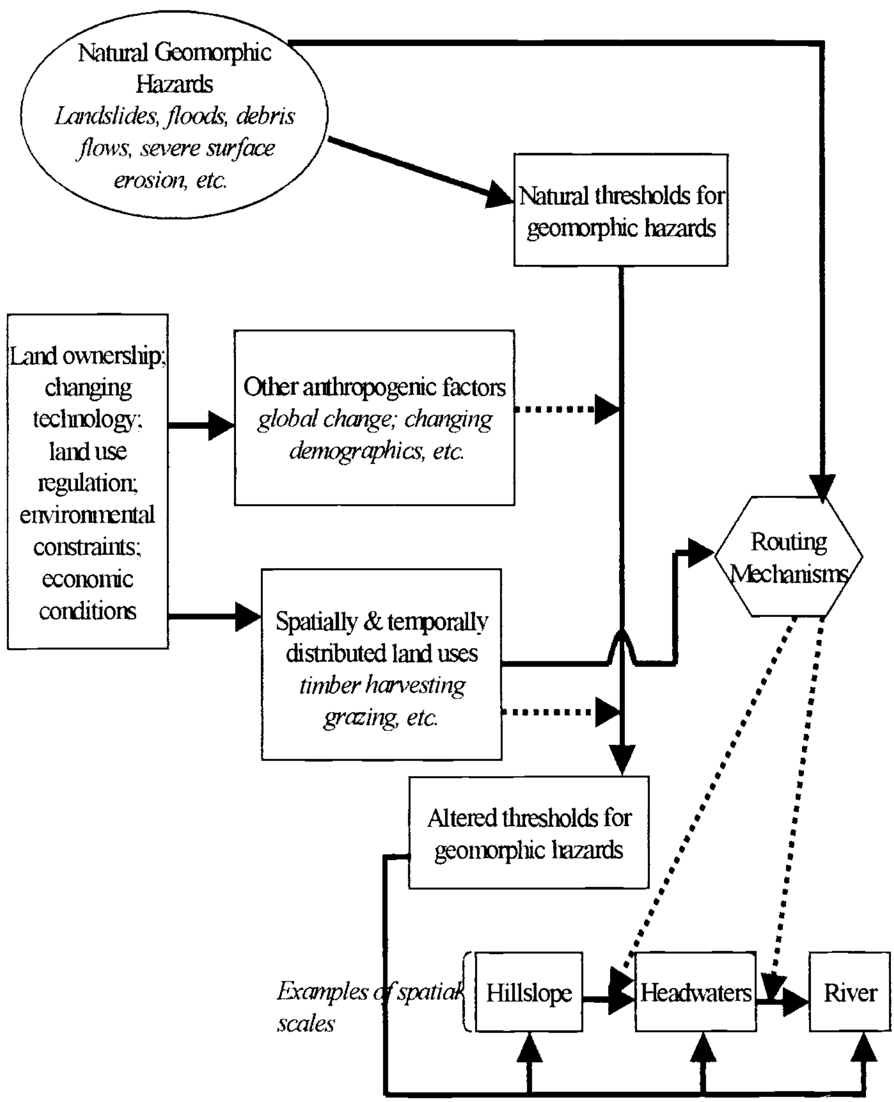


Fig. 11.1. A conceptual representation of the cumulative effects of anthropogenic factors on geomorphic hazards at various scales.

equivalent (Armstrong and Brodzik, 2001) can be estimated. However, dense vegetation canopies block ground images and attenuate radar signals (Tait, 1998; Wang *et al.*, 2000), thereby greatly negating the current utility of these methods in real-time hazard forecasting. The availability of more and more accurate DEMs has enhanced their utility in assessment of geomorphic hazards. Current applications of DEMs in geomorphic hazard assessment in forested regions include landslide (Montgomery and Dietrich, 1994; Wu and Sidle, 1995) and flood analysis (Wigmosta *et al.*, 1994). Other geomorphic applications are apparent, but these will undoubtedly benefit from higher resolution DEMs such as can be obtained from side-looking airborne radar and airborne laser altimetry (Dunne, 1998; Yang *et al.*, 2001). Together with these advanced remotely sensed data is the need to develop proper calibration: this requires field measurements. Thus, to progress, a concerted effort by government agencies, industrial and large private landowners and researchers is needed to collect and archive environmental data and to take advantage of developing spatial data handling and processing techniques in a way that will allow potential users reasonable access to this information (Dunne, 1998).

Significant advances in modelling related to geomorphic hazards have been made in the past decade. Distributed models allow us to simulate specific sites or at least regions where hazards may occur or propagate. However, a major current limitation is that the computational power of the many models has outstripped the resolution of input data, especially from remotely sensed sources. As better remotely sensed data become available, this problem will be somewhat nullified, however, major gaps will still persist and, in certain applications, only actual field data collected at detailed scales will suffice. Although attractive from the perspectives of ecosystem function as well as universal application, process-based models often do not accurately capture the underlying system processes. And inclusion of process details may be unrealistic both in terms of available data and computational demands, depending on scale. These problems are not intractable, but will require significant advancements in distributed data acquisition in areas with interfering canopy cover. In the meantime, the usefulness of empirical geomorphic hazard analyses that relay on spatially organized natural resource data should not be overlooked. Such hazard assessments have benefited greatly from advances in geographic information systems (GIS) along with remotely sensed data.

Future Challenges and Needs

Given current limitations in modelling geomorphic hazards, particularly with regard to availability of distributed data, one area that appears promising is the coupling of simple process-based models with digital elevation and other remotely sensed data. Such types of process-based, distributed models have already been developed for landslides and debris flow runout in managed forest terrain, although they have only been tested in a few areas (Wu and Sidle, 1995). Both process and semi-process based, distributed models have been used to predict surface erosion throughout the world (Beasley *et al.*, 1980; Young *et al.*, 1989), although most of these focus on estimating 'average' rates of erosion rather than episodic rates. Additionally, difficulties have been encountered in adapting empirical components of these models, based largely on extensive agricultural plot investigations, to more complex forested terrain. Distributed hydrological models have also been widely applied in situations with and without snowmelt (Wigmosta *et al.*, 1994), although parameter calibration is necessary to achieve accurate peak flow simulations. Overall, few of these distributed hydrologic and geomorphic models have been used to simulate the effects of possible climate change. Such applications together with incorporation of more explicit controlling processes and better attention to triggering thresholds will advance the state of geomorphic hazard modelling in forest terrain. Additionally, coupling field monitoring with modeling predictions holds promise for improving real-time hazard warning systems. Such a real-time warning system incorporating a high density of recording rain gauges has been implemented in the San Francisco Bay region to predict occurrence of debris flows (Keefer *et al.*, 1987).

Because many of the geomorphic hazards in primarily forested lands have a larger impact on natural resources than on humans and buildings, relatively little attention has been paid to hazard warning systems in this volume. Nevertheless, the vulnerability of humans, infrastructures and property to certain geomorphic hazards on forest lands cannot be ignored. In particular, coastal flooding due to typhoons and hurricanes advancing from the sea has devastated low-lying coastal communities (Ludlum, 1988; Inamura and Van To, 1997; Dube *et al.*, 2000). Also, landslides and related debris flows have claimed many lives and destroyed significant property in forested terrain (Sidle *et al.*, 1985; Swanston and Schuster, 1989). Downstream flooding has often been blamed on upper catchment forest harvesting, either rightly or wrongly (Douglas *et al.*, 1999; Lu *et al.*, 2001). Snow avalanches and icefall hazards have claimed extensive lives and inflicted considerable property damage in a few cases (Gurer *et al.*, 1995; Richardson and Reynolds, 2000). Volcanic hazards and earthquakes, while not specific in nature to forested terrain, have impacted many people due largely to the unpredictable nature of their occurrence (e.g., INCEDE, 1995; Shoaf *et al.*, 1998). For all of these hazards that occur with little notice, the implementation of appropriate warning systems can save lives and property. While such warning systems are not required in remote terrain, the increasing interface of urban centres (as well as the environmental pressures they bring to bear) and forests exemplifies the need for such real-time alerts. Unfortunately, developing nations, where hazard warning systems are most needed, are the most problematic in terms of implementation, due to lack of national coordination strategies, technology transfer gaps, and inability to pay for the service. Advances in geomorphic hazard warning systems, particularly in the developing world, will undoubtedly save lives and property and should be encouraged by national governments and international donors. Such a strategy should reap longer-term and more environmentally sustainable benefits compared to structural control of hazards. One cautionary pitfall that should be realized in planning and implementing hazard warning systems is that they must not perpetuate a false sense of security: that is to encourage people to return to live in high hazard areas because a hazard warning system has helped them avoid death or injury (Sorensen, 2000).

Geomorphic hazards that are more likely to be affected by climate change than by changes in forest land use also need further investigation. Such hazards are those more directly influenced by regional anticipated warming trends, for example permafrost, glacial, snow avalanche, weathering (related to rock failures) and snowmelt flood hazards. Typically these hazards occur at high latitudes (polar or sub-polar) where climate warming is projected to be greatest (Beniston, 2000). Process studies (particularly energy balance) need to be linked with modelling investigations to elucidate the spatial and temporal nature of hazard response, including better identification of triggering thresholds (Hoelzle *et al.*, 2001). While such hazards will not directly impact large numbers of people, they can have long-term consequences for the extent and ecosystem processes in boreal forests and tundra.

The effects of changing forest environments on geomorphic hazards will be manifested most strongly in the developing tropics. These regions not only are naturally vulnerable to a wide array of hazards, such as coastal flooding, landslides, debris and pyroclastic flows, riverine flooding, tsunami, severe surface erosion and salinization, they are also most affected by socio-political pressures and related changing patterns of land use. Advances in hazard assessment in these regions will need to include socio-economic dimensions as well as process-based scientific inputs. In all environments, it is the interaction among human-induced disturbances, natural geomorphic and ecosystem processes, and climate that determines how climate change will be manifested in particular settings.

References

Armstrong, R.L. and Brodzik, M.J. (2001) Recent Northern Hemisphere snow extent: a comparison of data derived from visible and microwave satellite sensors. *Geophysical Research Letters* 28, 3673-3676.

Beasley, D.B., Huggins, L.F. and Monke, E.J. (1980) ANSWERS: a model for watershed planning. *Transactions, American Society of Agricultural Engineers* 23, 938-944.

Beniston, M. (2000) *Environmental change in mountains and uplands. Key Issues in Environmental Change.* Edward Arnold, London, and Oxford University Press, Oxford.

Bevers, M., Hof, J. and Troendle, C. (1996) Spatially optimizing forest management schedules to meet stormflow constraints. *Water Resources Bulletin* 32, 1007-1015.

Douglas, I., Bidin, K., Balamurugan, G., Chappell, N.A., Walsh, R.P.D., Greer, T. and Sinun, W. (1999) The role of extreme events in the impacts of selective tropical forestry on erosion during harvesting and recovery phases at Danum Valley, Sabah. *Philosophical Transactions, Royal Society of London, B* 354, 1749-1761.

Dube, S.K., Chittibabu, P., Rao, A.D., Sinha, P.C. and Murty, T.S. (2000) Extreme sea levels associated with severe tropical cyclones hitting Orissa Coast of India. *Marine Geodesy* 23, 75-90.

Dunne, T. (1998) Critical data requirements for prediction of erosion and sedimentation in mountain drainage basins. *Journal of the American Water Resources Association* 34, 795-808.

Gurer, I., Tuncel, H., Murat Yavas, O., Erenbilge, T. and Sayin, A. (1995) Snow avalanche incidents in north-western Anatolia, Turkey during December 1992. *Natural Hazards* 11, 1-16.

Hoelzle, M., Mittaz, C., Etzelmüller, B. and Haeberli, W. (2001) Surface energy fluxes and distribution models of permafrost in European mountain areas: an overview of current developments. *Permafrost and Periglacial Processes* 12, 53-68.

Hugh, E. and Lambin, E.F. (2000) Fires and land-cover change in the tropics: a remote sensing analysis at the landscape scale. *Journal of Biogeography* 27, 765-776.

Inamura, F. and Van To, D. (1997) Flood and typhoon disasters in Viet Nam in the half century since 1950. *Natural Hazards* 15, 71-87.

INCEDE (International Center for Disaster-Mitigation Engineering) (1995) The first 55 hours – Great Hanshin Earthquake, January 17, 1995. *INCEDE Newsletter (Special Issue, January 1995)*, Institute of Industrial Science, The University of Tokyo, Japan.

Keefer, D.K., Wilson, R.C., Mark, R.K., Brabb, E.E., Brown, W.M., Ellen, S.D., Harp, E.L., Wieczorek, G.F., Alger, C.S. and Zatkin, R.S. (1987) Real-time landslide warning during heavy rainfall. *Science* 238, 921-925.

Kiely, G. (1999) Climate change in Ireland from precipitation and streamflow observations. *Advances in Water Resources* 23, 141-151.

Lu, S.-Y., Cheng, J.D. and Brooks, K.N. (2001) Managing forests for watershed protection in Taiwan. *Forest Ecology and Management* 143, 77-85.

Ludlum, D.M. (1988) The great hurricane of 1938 (eastern USA). *Weatherwise* 41, 214-216.

Marzano, F.S., Turk, J.F., Ciotti, P., DiMichele, S. and Pierdicca, N. (2001) Potential of combined spaceborne infrared and microwave radiometry for near real-time rainfall attenuation monitoring along Earth-satellite links. *International Journal of Satellite Communications* 19, 385-412.

McGurk, B.J. and Fong, D.R. (1995) Equivalent roaded area as a measure of cumulative effect of logging. *Environmental Management* 19, 609-621.

Montgomery, D.R. and Dietrich, W.E. (1994) A physically based model for the topographic control on shallow landsliding. *Water Resources Research* 30, 1153-1171.

Moran, M.S., Hymer, D.C., Qi, J. and Sano, E.E. (2000) Soil moisture evaluation using multi-temporal synthetic aperture radar (SAR) in semiarid rangeland. *Agricultural and Forest Meteorology* 105, 69-80.

Pickup, G. and Marks, A. (2001) Regional-scale sedimentation process models from airborne gamma ray remote sensing and digital elevation data. *Earth Surface Processes and Landforms* 26, 273-293.

Richardson, S.D. and Reynolds, J.M. (2000) An overview of glacial hazards in the Himalayas. *Quaternary International* 65-66, 31-47.

Shoaf, K.I., Sareen, H.R., Nguyen, L.H. and Bourque, L.B. (1998) Injuries as a result of California earthquakes in the past decade. *Disasters* 22, 218-235.

Sidle, R.C. and Hornbeck, J.W. (1991) Cumulative effects: a broader approach to water quality research. *Journal of Soil and Water Conservation* 46, 268-271.

Sidle, R.C. and Wu, W. (1999) Simulating the effect of timber harvesting on temporal and spatial distribution of shallow landslides. *Zeitschrift für Geomorphologie* 43, 185-201.

Sidle, R.C., Pearce, A.J. and O'Loughlin, C.L. (1985) *Hillslope stability and land use.* Water Resources Monograph, No. 11, American Geophysical Union, Washington, DC.

Sorensen, J.H. (2000) Hazard warning systems: review of 20 years of progress. *Natural Hazards Review* 1, 119-125.

Swanston, D.N. and Schuster, R.L. (1989) Long-term landslide hazard mitigation programs: structure and experience from other countries. *Bulletin of the Association of Engineering Geologists* 26, 109-133.

Tait, A.B. (1998) Estimation of snow water equivalent using passive microwave radiation data. *Remote Sensing of Environment* 64, 286-291.

Thornton, K.W., Holbrook, S.P., Stolte, K.L. and Landy, R.B. (2000) Effects of forest management practices on mid-Atlantic streams. *Environmental Monitoring and Assessment* 63, 31-41.

Wang, Y., Kasischke, E.S., Bourgeau-Chavez, L.L., O'Neill, K.P. and French, N.H.F. (2000) Assessing the influence of vegetation cover on soil-moisture signatures in fire-disturbed boreal forests in interior Alaska: modeled results. *International Journal of Remote Sensing* 21, 689-708.

Wigmosta, M.S., Vail, L.W. and Lettenmaier, D.P. (1994) A distributed hydrology-vegetation model for complex terrain. *Water Resources Research* 30, 1665-1679.

Wu, W. and Sidle, R.C. (1995) A distributed slope stability model for steep forested hillslopes. *Water Resources Research* 31, 2097-2110.

Yang, D., Herath, S. and Musiake, K. (2001) Spatial resolution sensitivity of catchment geomorphologic properties and the effect on hydrological simulation. *Hydrological Processes* 15, 2085-2099.

Young, R.A., Onstad, C.A., Bosch, D.D. and Anderson, W.E. (1989) AGNPS: A nonpoint-source pollution model for evaluating agriculture watersheds. *Journal of Soil and Water Conservation* 44, 168-173.

Index