Local scale assessment of climate change and its impacts in the Highlands and Islands of Scotland

Thesis

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'Local scale assessment of climate change and its impacts in the Highlands and Islands of Scotland'©

By
John Coll

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UHI Millennium Institute

The Open University

Environmental Research Institute,
The North Highland College,
UHI Millennium Institute,
Thurso KW14 7JD, Scotland

The Open University
Milton Keynes,
MK7 6AA, UK

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Abstract

The global climate is warming and there is consensus that recent warming trends will amplify, as the present century progresses, in response to a continued build up of atmospheric greenhouse gas (GHG) concentrations. However, there are limitations associated with Global Climate Model (GCM) and Regional Climate Model (RCM) outputs for topographically diverse regions. Strategic management decisions relating to maritime upland communities require locally resolved projections of change across a range of elevations, which are not supplied by the present generation of models.

Here, some of these challenges are addressed via a series of distinctive analyses. Quality controlled baseline station data are used to assess performance outputs for seasonal mean values of temperature and precipitation from an RCM at representative locations across the region. In the case of temperature these inter-comparisons indicate a warm bias in the RCM-simulated seasonal minima for the transition seasons of spring and autumn, whereas for summer maxima there is a cold bias in RCM-simulated values.

RCM-generated outputs of future changes to temperature and precipitation are then variably combined with station data to model altitudinal changes at western and eastern upland locations. These analyses indicate a substantial upward migration in key seasonal temperature isotherms associated with present vegetation zones for the climate change scenarios used. This approach is then extended by applying selected outputs to conduct Climate Change Impact Assessments (CCIAs) for the scenarios used in a series of upland case studies.

The analyses flag a number of remaining research challenges. Principally, these are that scale-dependent controls on local topo-climates are not adequately captured in the GCM driven RCM projection. While the approach delivers more refined local-scale projections of possible change across a range of elevations than has hitherto been available, residual uncertainties associated with the use of GCM/RCM outputs remain.

Glossary

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Glossary

Alphabetical List of Acronyms

AO	Arctic Oscillation
AOCGCM	Atmosphere-Ocean Coupled Global Climate Model
AWS	Automated Weather Station
BADC	British Atmospheric Data Centre
CCIAs	Climate Change Impact Assessments
CRU	Climatic Research Unit, University of East Anglia
EA	United Kingdom Environment Agency
ENSO	El Nino Southern Oscillation
CGM(s)	Global Climate Model(s)
GHGs	Greenhouse Gases
HadAM3H	Hadley Centre intermediate scale atmosphere only climate model
HadAM3P	Hadley Centre intermediate scale atmosphere only climate model (new version)
HadCM2	Hadley Centre ocean-atmosphere coupled GCM, version 2
HadCM3	Hadley Centre ocean-atmosphere coupled GCM, version 3
HadGEM1	Hadley Centre Global Environmental Model, version 1
HadRM2	Hadley Centre Regional Climate Model, version 2
HadRM3	Hadley Centre Regional Climate Model, version 3
IPCC	Intergovernmental Panel on Climate Change
IPCC-SRES	Intergovernmental Panel on Climate Change Special Report on Emissions Scenarios
IPCC-TAR	Intergovernmental Panel on Climate Change Third Assessment Report
IPCC-TGCIA	Intergovernmental Panel on Climate Change Technical Guidance Committee on Climate Change Impact Assessment
LRM(s)	Lapse Rate Model(s)
LV	Lapse Value
Met Office	United Kingdom Meteorological Office
NAO	North Atlantic Oscillation
NAOI	North Atlantic Oscillation Index

OST	United Kingdom Office of Science and Technology
Precip	Mean total precipitation
RCM(s)	Regional Climate Model(s)
SAT	Surface Air Temperature
SNIFFER	Scottish and Northern Ireland Forum for Environmental Research
THC	Thermohaline Circulation
T_{max}	Mean maximum temperature
T_{mean}	Mean temperature
T_{min}	Mean minimum temperature
UKCIP	United Kingdom Climate Impacts Programme
UKCIP98	United Kingdom Climate Impacts Programme Climate Change Scenarios 1998
UKCIP02	United Kingdom Climate Impacts Programme Climate Change Scenarios 2002
UKCIP-Next	United Kingdom Climate Impacts Programme Climate Change 'Next' Scenarios (~ spring 2008)
UKMO	United Kingdom Meteorological Office
WMO	World Meteorological Association

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My very special thanks and love must of course go to Trish, Luke and Catherine for their support and for putting up with me this past year. Not least for tolerating the strange comings and goings at all sorts of unusual hours during the writing process.

This research is dedicated to my parents, without them there would be no writing. My pain and grief at the loss of my dad halfway through this project was, and is, but a fraction of that endured by my mum in losing her lifelong partner. This book is committed with love to his memory and in the hope that any sense of pride my mum may take in it will in some small way ease her burden.

1. Introduction

1.1 A Wider Context for Regional Climate Change Assessment

1.1.1 Drivers of recent global and hemispheric trends

Since the beginning of the industrial revolution carbon dioxide (CO_2) in the atmosphere has increased to a level that has not been exceeded over the last 420,000 years, and probably not during the past 20 million years (IPCC, 2001; WMO, 2006). This steady rise in atmospheric CO_2 levels has seen an increase of ~100 parts per million (ppm), from ~280 ppm at the beginning of industrialisation (Monnin *et al.*, 2001; Ruttiman, 2006) to the present level of ~380 ppm (Ruttiman, 2006). These increases have resulted primarily from the burning of fossil fuels, with additional contributions from deforestation and changes in land management practises (Houghton, 2002).

With CO_2 concentrations projected to increase from the current level of ~380 ppm (Ruttiman, 2006) to between 540 ppm and 1000 ppm over the next 100 years (depending on assumptions regarding future greenhouse gas emissions), temperatures are predicted to rise 1.4 to 5.8°C above present levels (Houghton, 2002). In the most recent Hadley Centre Global Environmental Model (HadGEM1) simulation, the global mean surface temperature rise for 2071-2100 is projected at ~3.4 K to ~3.8 K relative to 1961-1990 for the Special Report on Emissions Scenarios (SRES) used (Stott *et al.*, 2006).

Already, globally averaged concentrations of CO_2, methane (CH_4) and nitrous oxide (N_2O) in the Earth's atmosphere have reached their highest-ever recorded levels in 2004, when CO_2 was recorded at 377.1 ppm, CH_4 at 1783 parts per billion (ppb), and

N_2O at 318.6 ppb (WMO, 2006). These supersede pre-industrial values by 35%, 155% and 18% respectively, an increase over the previous decade by 19ppm, 37ppb and 8ppb in absolute amounts (WMO, 2006). There is good evidence that higher global temperatures will promote a further rise of greenhouse gas (GHG) levels, implying a positive feedback which will increase the the effect of anthropogenic emissions on global temperatures (Scheffer *at al.*, 2006; Torn and Harte, 2006).

1.1.2 Sources of variability in the climate system

In addition to GHGs, there are a number of sources of natural variability in surface air temperatures, notably those due to internal interactions within the climate system, such as the El Nino Southern Oscillation (ENSO) and external forcing through, for example, changes in solar luminosity or volcanic activity (Barnett *et al.*, 1999). However, the magnitude of natural climate forcings is relatively uncertain, instrumental observations of variations in solar irradiance are available only for the last two solar cycles (~22 years) and estimates of historical solar forcing must therefore rely on reconstructions from proxy sources (Zwiers and Weaver, 2000).

While these reconstructions differ widely, they concur that the forcing is positive over the 20th century and that it has increased by around 10 to 20% of the change in greenhouse gas forcing (Zwiers and Weaver, 2000). Reconstructions of volcanic forcing are similarly uncertain because instrumental observations of stratospheric aerosol loading are available only for the past two decades and there is a similar reliance on phenomenological and proxy observations before that time (Zwiers and Weaver, 2000).

The global mean average surface temperature has risen in the last two decades and has reached its highest level during the last 140 years – the period of the instrumental record (IPCC, 2001; Easterling *et al.*, 2000). Trends in mean annual air temperature for 1950-98 indicate three areas of particularly rapid regional warming, all at high latitudes (Hansen *et al.*, 1999; Vaughan *et al.*, 2004): north-western North America and the Beaufort Sea, an area around the Siberian Plateau, and the Antarctic Peninsula and Bellingshausen Sea (Vaughan *et al.*, 2004) (Figure 1.1). Figure 1.1 also usefully illustrates the spatial and temporal variability in the pattern of warming and indicates that over the past two decades, the warming has become global (Jones and Moberg, 2003).

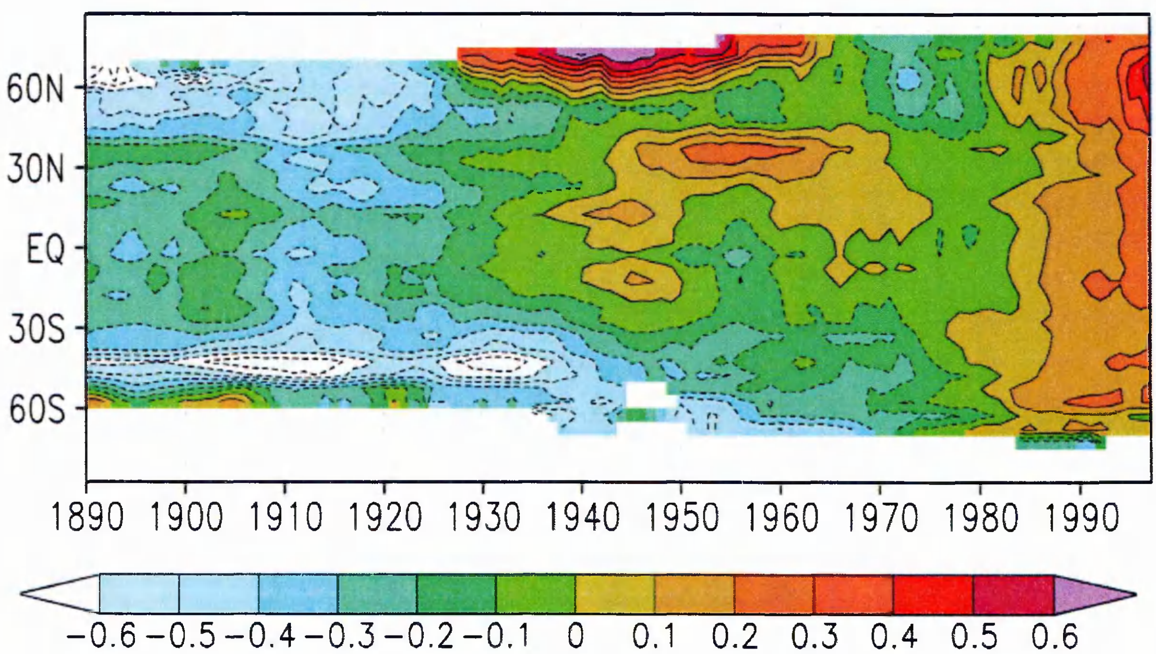


Figure 1.1: Observed global surface air temperature (SAT) anomaly plot (°C) by latitude and time 1890-2000 indicating the temporal and spatial variation of warming episodes (Source: Morison, 2003).

Analyses of data from the northern hemisphere show that the increase in temperature during the 20th century is likely to be the largest of any century during the last 1000 years (Houghton, 2002; Osborn and Briffa, 2006) (Figure 1.2). It seems probable, therefore, that anthropogenic forcing of climate can best explain recent anomalies (Jones and Mann, 2004; Knutson *et al.*, 2006; Stott *et al.*, 2006). For further information the reader is directed to Sections 3.4.1 and 3.4.2 where some of these wider trends are linked to aspects of variability in the Scottish data used in this work.

1.1.3 Detection and attribution of recent changes

In order to have more confidence in model simulations of future climate, it is useful to detect and quantify the model simulated anthropogenic climate change in observations. Many studies have shown that the most recent observations of changes in surface and free-atmosphere temperatures cannot be explained by (model-estimated) natural climate variability alone (Hegerl *et al.*, 2000; Barnett *et al.*, 2005). Thus approaches have been developed which rely on data from climate models to estimate both the space-time pattern of the climate response to external forcings (climate change signals) and the magnitude and patterns of internal climate variability (Hegerl, *et al.*, 2000; Stott *et al.*, 2000, 2002, 2006; Knutson *et al.*, 2006).

Tett *et al.* (1999, 2000, 2002) considered that solar forcing may have contributed to the temperature changes early in the century, with anthropogenic causes combined with natural variability also presenting a possible explanation. Further studies also suggest that, regardless of any possible amplification of solar or volcanic influence, purely natural forcing could be excluded and the warming over the latter part of the century could be largely attributed to anthropogenic components (Stott *et al.*, 2000; Stott, 2003; Huntingford *et al.*, 2006; Knutson *et al.*, 2006; Stott *et al.*, 2006). And that, in particular, warming in the trophosphere since the 1960s is probably mainly due to anthropogenic forcings with a negligible contribution from natural forcings

(Tett *et al.*, 2000, 2002). While Houghton *et al.* (2001) in the IPCC Third Assessment Report (TAR) concluded that most of the observed warming over the last 50 years is likely to be due to the increased concentration of GHGs. Overall, it is considered that detection studies show that anthropogenic climate change is detectable in the surface temperature records of individual continents and that it can be distinguished from climate change because of natural forcing (Barnett *et al.*, 2005; Knutson *et al.*, 2006; Stott *et al.*, 2006).

The extended analysis of Stott *et al.* (2001) and Stott (2003) broadly supports these conclusions, but with the caveat that the evidence for the importance of natural effects for the early century warming remains less clear-cut. Despite employing different methods which took account of model errors, errors in the forcing and errors in the observations, Hegerl *et al.* (2000) reached broadly similar conclusions, i.e. there is suggestive evidence that a combination of natural and anthropogenic forcings has influenced the twentieth century climate evolution. A conclusion broadly supported by Thorne *et al.*, (2002) in assessing the robustness of previous optimal detection and attribution studies considering zonal-mean temperatures. The early century warming was a very unusual event in a 1000 year proxy record (Figure 1.2), with results indicating that it was caused either by a superposition of anthropogenic climate change and a very extreme event of internal climate variability, or is influenced by additional natural forcing.

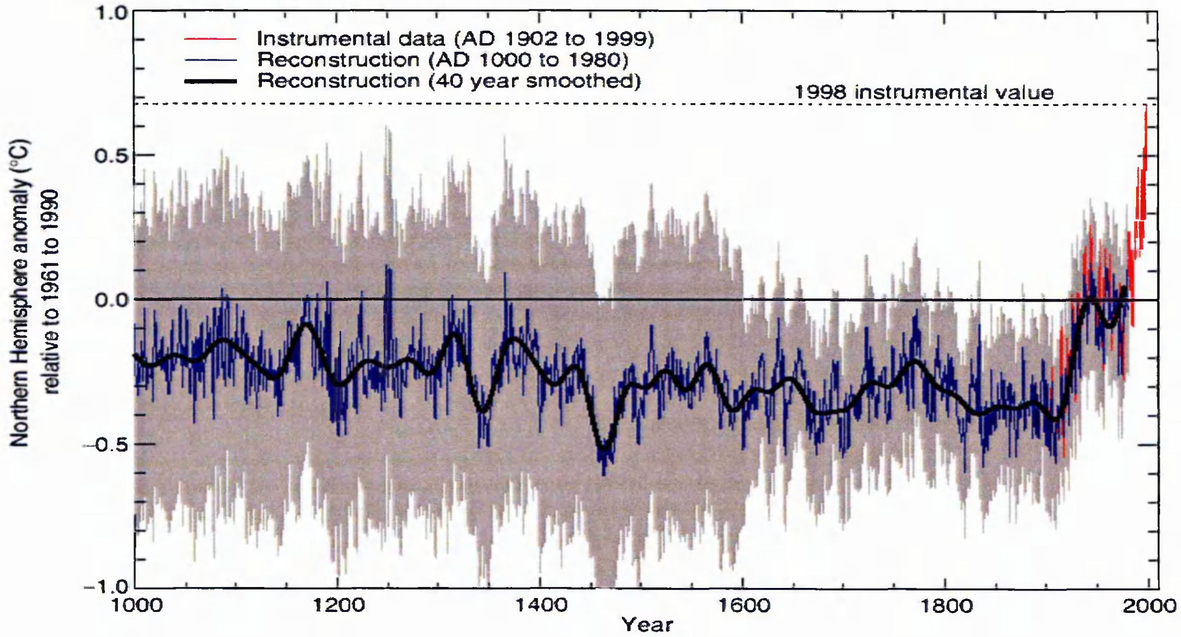


Figure 1.2: Northern Hemisphere average annual SAT variations over the last millennium from proxy, historical and instrumental observations. Based on the analysis of Mann *et al.*, 1998; 1999 (Source IPCC; 2001).

However, these analyses did not include the effects of other forcings such as changes in land-surface properties and mineral dust which could have affected climate (Tett *et al.*, 2000; Stott *et al.*, 2001). Easterling *et al.* (2000) were similarly of the view that the qualitative consistency among the observations over the 20th century, and the models for the end of the 21st century, suggests that at least some of the changes observed to date are likely associated with changes in forcing that we have already experienced over the 20th century.

By contrast other workers have been more equivocal in their interpretation urging caution in the relative ascription of warming, citing factors such as:

- With some notable exceptions, local instrumental records are generally less than 100 years in length;

- the instrumental record is likely to be contaminated with an anthropogenic signal if one is present and attempts at anthropogenic signal removal leave a residual that can be attributed to natural variability;
- biases in the instrumental record due to changing measurement methods/instruments allied to the fact that there are large expanses of the planet where observations are either scant or missing, particularly for the early part of the 20th century (Barnett *et al.*, 1999).

Consequently, they concluded that the most probable cause of the observed warming is a combination of internally and externally forced natural variability and anthropogenic sources. Therefore the observed changes in global and large-scale regional climate could not be attributed to anthropogenic forcing alone. In any event the global-mean surface air temperature was broken four times during the 1990s (1990, 1995, 1997 and 1998) ensuring it was the warmest decade since 1860 when the global-mean surface air temperature record commenced and possibly also the warmest decade this millennium for the Northern Hemisphere (Hulme, 1999; Brohan *et al.*, 2006). These trends have continued into the new millennium with four of the five warmest years in the Northern Hemisphere record occurring since 2001, with 2003 as the warmest year in the record (CRU, 2005). The year 2005 was equal second warmest on record globally, exceeded only by 1998 (Brohan *et al.*, 2006).

The associated unusually hot and dry summer of 2003 had a huge impact on biomass productivity in Europe, with more than a year's recovery time having been estimated (Ciais *et al.*, 2005). While analysis of a 60 year temperature record for the North

Atlantic region suggests the 2003 summer event was part of an increase in the recent frequency of extreme warm events (Nogaj *et al.*, 2006). Against this backdrop, there is a growing expectation that increases in the concentration of GHGs arising from human activity will lead to substantial changes in climate in the twenty first century (Johns *et al.*, 2003). Although, as yet there is no single scientific assessment that can be made of a warming commitment based on emissions scenarios (Hare and Meinhausen, 2006).

Paleo-climatic data indicate that this recent rate of increase is likely to be without precedent during at least the last 10,000 years (Albritton *et al.*, 2001; Jones and Mann, 2004). While paleo-reconstructions of Northern Hemispheric mean temperatures show that the 20th century warming is unique in the last millennium both for its size and rapidity (Jones and Moberg, 2003; von Storch *et al.*, 2004; Barnett *et al.*, 2005). (Figure 1.2).

However, recent work has questioned the validity of the iconic 'hockey stick' graph (Figure 1.2), with the scientific debate focussed on the validity and rigour of the methods used to reconstruct temperatures in the 15th century (McIntyre and McKitrick, 2005a, b). This has led other workers (von Storch and Zorita, 2005) to defend the original work (Mann *et al.*, 1998; 1999) based on their own more recent approach to reconstructing paleo-temperature records (von Storch *et al.*, 2004). With the scientific debate seized upon by contrarian senators in the US Congress as a refutation of global warming, a sub-committee of the US National Academies has recently (2006) moved to defend the reconstruction, although with some qualifications on why there is less confidence in temperature reconstructions from A.D. 900 to 1600.

1.1.4 Projected future changes at global and hemispheric scales

Associated with projected future warming, landmasses are expected to warm more than the oceans and high latitudes more than the tropics (Allbritton *et al.*, 2001: Giorgi, 2005). This polar amplification and a greater warming of the land compared to the oceans is also a feature in the latest HadGEM1 simulation (Stott *et al.*, 2006). Although less certain, projections from models generally show increased precipitation in the high latitudes of both hemispheres (Palmer and Raisanen, 2002; Frei *et al.*, 2003; Kundzewickz *et al.*, 2006), and in the tropics (Houghton, 2002). These simulated precipitation increases from mid-latitudes to the Polar regions can best be explained by the increased ability of the warmed atmosphere to more effectively transport moist air polewards (Christensen, 2001).

Most of the models also simulate increases in soil moisture in northern areas in winter due to more precipitation falling as rain rather than snow in a warmed atmosphere, whereas in summer there seems to be a general tendency towards mid-latitude soil drying in the simulations (Christensen, 2001). Despite these possible reductions in average summer precipitation over many parts of the European continent, precipitation amounts exceeding the 95th percentile are very likely in many areas, thus episodes of severe flooding may become more frequent despite the general trend towards drier summer conditions (Christensen and Christensen, 2002).

During the late boreal summer (July-September), European climatic conditions occasionally favour very severe precipitation episodes, such as those that caused the recent flooding of the rivers Odra (1997), the Elbe and sub-catchments of the Rhone in 2002 (Christensen and Christensen, 2002). This may be explained by the fact that the atmosphere will contain more water in a warmer climate, which will provide further potential for latent-heat release during the build-up of low-pressure systems, thereby possibly both intensifying the systems and making more water available for precipitation (Christensen and Christensen, 2002).

In general Global Climate Models (GCMs) simulate the greatest future warming over the northern part of the globe (Giorgi, 2006), with the greatest increases in autumn and winter a consequence of sea ice being formed later and in lower quantities in a warmer climate (Christensen, 2001). However there are discrepancies between model simulated changes to sea ice distribution, and hence the magnitude of future warming depending on how the different models treat the physical processes involved in the formation and destruction of sea-ice e.g. (Christensen, 2001). Some of the limitations surrounding current regional climate modelling are dealt with further in the Introduction to Chapter 4, while the GCMs used for the forthcoming IPCC 4th Assessment Report (AR4) are summarised in Table 1.1.

Table 1.1 GCMs used for the forthcoming IPCC Assessment Report 4 (AR4), their originating groups and country

Model	Group	Country
CCSM3	National Center for Atmospheric Research	USA
CGCM3.1 (T42)	Canadian Centre for Climate Modelling	Canada
CGCM3.1 (T63)	Canadian Centre for Climate Modelling	Canada
CNRM-CM3	Centre National de Recherches Meteorologiques	France
CSIRO-Mk3.0	CSIRO Atmospheric Research	Australia
ECHAM5/MPI-OM	Max Planck Institute for Meteorology	Germany
FGOALS-g1.0	LASG/Institute of Atmospheric Physics	China
GISS-AOM	National Center for Atmospheric Research	USA
INM-CM3.0	Institute for Numerical Mathematics	Russia
IPSL-CM4	Institut Pierre Simon Laplace	France

MRI-CGCM2.3.2	Meteorological Research Institute	Japan
PCM	National Center for Atmospheric Research	USA
UKMO-HadCM3	Hadley Centre for Climate Prediction and Research	UK

Despite these issues surrounding the detail of outputs from different models, NW Europe and high latitude areas continue to be identified as regional climate change 'Hot-Spots' (*sensu* Giorgi, 2006). In an ensemble-based approach using outputs from 20 AOCGCMs for three IPCC emission scenarios, after the Mediterranean and North Eastern European regions, NW Europe is also identified as a warming hot spot (Giorgi, 2006). This shift towards the ability to produce probabilistic predictions of climate change for use in impact assessment work (Murphy, 2004; Giorgi, 2005) based on ever increasing computer power is briefly discussed further in Chapter 8.

1.2 Hemispheric and Regional Influences on NW European Climate

1.2.1 North Atlantic climate stability – the role of the Atlantic thermohaline (THC) circulation

Ocean currents cause significant geographical differences in the supply of heat to to the atmosphere and regions around the North Atlantic Ocean have a mean annual surface air temperature that is 5–7°C warmer than those at the same latitude in the Pacific (Stocker and Marchal, 2000). This can be attributed to the thermohaline circulation (THC) of the Atlantic Ocean moving warm, saline tropical waters northward, the Gulf Stream being part of this basin-scale circulation (Stocker and Marchal, 2000) (Figure 1.3). The deep convection and sinking in the North Atlantic have no counterpart in the North Pacific Ocean, where northward heat transport is much weaker (Srokosz and Gommenginger, 2002).

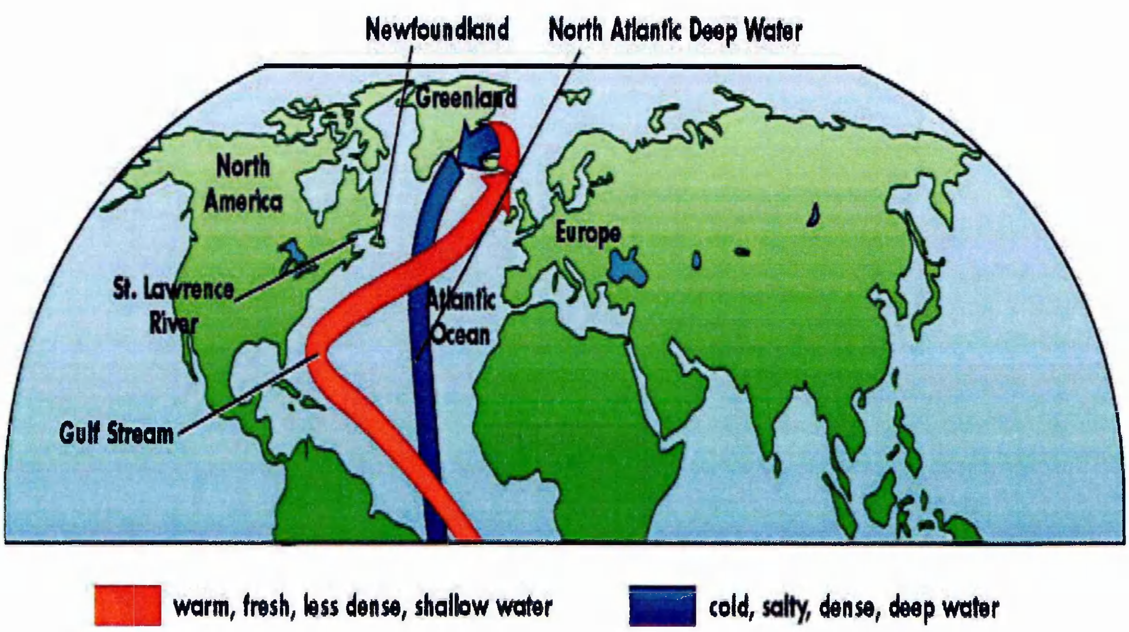


Figure 1.3: Simplified cartoon representation of the North Atlantic component of the thermohaline circulation (THC) (Source: NASA GSFC).

These synoptic eddies are analogous to the high and low-pressure systems that make up the weather in the atmosphere and play an important role in mixing ocean waters and transporting heat between regions (Rahmstorf, 1999; Dai *et al.*, 2005; Stouffer *et al.*, 2006). Sometimes popularly dubbed the 'conveyor belt' this large scale overturning motion of the Atlantic drives warm surface waters northwards, with cold deep water returning southwards (Rahmstorf, 1999; Srokosz and Gommenginger, 2002) and largely determines the mild climate at the west-side of the European continent (Klein Tank and Konnen, 1997; Srokosz and Gommenginger, 2002).

The existence of abrupt past climate changes has fuelled concern over the possibility of similar changes in future, particularly if anthropogenic climate change might trigger another instability of the circulation and a severe cooling over the North Atlantic and parts of Europe (Rahmstorf and Ganopolski, 1999; Rahmstorf, 2000; Srokosz and Gommenginger, 2002; Stouffer *et al.*, 2006). However, the driving forces for the frequent abrupt and severe climate fluctuations during the last glacial and Holocene remain unclear, while Atmosphere Ocean Coupled Global Climate Model (AOCGCM) experiments have only simulated such events by inserting large amounts of fresh water into the northern North Atlantic Ocean (Hall and Stouffer, 2001; Vellinga, 2004). Nonetheless, the ultimate forcing of millennial-scale climate change remains elusive, despite the bi-polar seesaw of temperature fluctuations being an apparently persistent feature of glacial climate (Stocker and Marchal, 2000; Blunier and Brook, 2001), and despite this north-south temperature anti-phase being captured in climate model studies (Stocker and Schmittner, 1997; Schiller *et al.*, 1997; Stocker and Marchal, 2000; Blunier and Brook, 2001). Although recent coupled

AOCGCM experiments have cautioned that the concept of an ocean bi-polar seesaw should be subject to some caveats (Seidov *et al.*, 2005).

Ocean models suggest that an increase in North Atlantic THC cools at least parts of the Southern Hemisphere and warms the high-latitude Northern Hemisphere, with this asynchrony appearing in a number of model runs (Blunier and Brook, 2001; Griesel and Maqueda, 2006). The THC is a self-sustaining phenomenon and is thus prone to instability, although it can also exhibit more than one stable equilibrium as well as hysteresis behaviour, a feature typical of a non-linear physical system (Stocker and Marchal, 2000) and a behaviour which is captured across models (Rahmstorf *et al.*, 2005).

Both the paleo-climatic record and modelling studies support the notion that the climate system has more than one equilibrium state, and that perturbations can trigger transitions between them (Stocker and Schmittner, 1997; Stouffer *et al.*, 2006). For example, various paleo-climatic indices reveal that both during the transition from the last ice age into the present interglacial and during the last interglacial, the climate in the North Atlantic region showed large and rapid changes (Johnsen *et al.*, 1992; Bond *et al.*, 1993; Dansgaard *et al.*, 1993; Taylor *et al.*, 1993; Field *et al.*, 1994; Keigwin *et al.*, 1994; Klein Tank and Konnen, 1997; Hall and Stouffer, 2001; Romanova *et al.*, 2006) with rapid oscillations between cold and warm states lasting for several thousand years throughout the last glacial (Blunier and Brook, 2001). However, these fluctuations are only part of climate system variability on a longer timescale, with the obvious unknown being how anthropogenic warming superimposed on natural variability may affect THC behaviour in the future.

With a warming earth, two factors affect the density of ocean waters and thereby the thermohaline circulation; the temperature increase and the change in freshwater budget (Rahmstorf, 2000). However, there is uncertainty (Section 1.4) associated with regional precipitation changes due to global warming (Rahmstorf, 1997; 1999) and the possible impacts of greater freshwater inputs from increased precipitation on surface water density affecting oceanic circulation. Of the two major locations of deep water formation, the Greenland-Iceland-Norwegian Sea and the Labrador Sea (Stocker and Marchal, 2000), it is thought that the Labrador convection site may be the more vulnerable of the two since the deep water formed in the Labrador Sea is already the less dense (Rahmstorf, 1999). While some models indicate that unchecked greenhouse gas emissions could trigger a circulation shutdown in the Atlantic (Rahmstorf and Ganopolski, 1999), the relative importance of temperature increase and freshwater input in weakening the Atlantic conveyor belt differs between different models (Dixon *et al.*, 1999; Rahmstorf and Ganopolski, 1999; Rahmstorf, 2000; Stouffer *et al.*, 2006). However, and perhaps of considerable significance, recent monitoring work at 25°N in the Atlantic indicates a THC slowing of 30% between 1957 and 2004 (Bryden *et al.*, 2005).

While the models generally agree that during the phase of GHG increase a weakening or even collapse of the thermohaline circulation does not lead to a surface cooling below pre-industrial levels, serious cooling of the North Atlantic region occurs only in the longer term when GHGs decline again and the circulation remains in the 'off' mode (Rahmstorf, 2000). However, despite the many qualitative and sometimes quantitative agreements, serious gaps and shortcomings remain in our ability to simulate abrupt climate change (Stocker and Marchal, 2000; Wunsch, 2006). Further uncertainty here arises from many of the key processes involving atmospheric heat and water transports which control the stability of the THC being subject to large modelling uncertainty (Rahmstorf and Ganopolski, 1999; Wood *et al.*, 1999; Latif *et al.*, 2000; Thorpe *et al.*, 2001; Vellinga *et al.*, 2001; Srokosz and Gommenginger 2002; Higgins and Vellinga, 2004; Vellinga, 2004; Schmittner *et al.*, 2005).

Despite these considerable uncertainties surrounding the stability and behaviour of the THC on varying timescales, consensus remains that a major ocean circulation change should be considered a 'low probability-high impact' risk, but emphasises that proper risk analysis is crucial for this type of non-linear climatic change (Rahmstorf, 2000; Hulme *et al.*, 2002; Schmittner *et al.*, 2005). Nonetheless there remains the possibility that THC changes could result in substantial and rapid climate change for western Europe and Scandinavia. However, at present levels of knowledge it is not possible to quantify the probability of this occurring (Srokosz and Gommenginger, 2002; Schmittner *et al.*, 2005; Hargreaves and Annan, 2006).

Given the complexity of climate system linkages, this is not surprising. Some AOCGCM work emphasises the role of Southern Ocean wind stress in affecting mixing characteristics (Griesel and Maqueda, 2006; Liu, 2006). While other work indicates a THC response in phase with the North Atlantic Oscillation (NAO) (Brauch and Gerdes, 2005: Belluci and Richards, 2006; Brauch), suggesting a dominant role for the advective mixing mechanism (Dai *et al.*, 2005). This has led some authors to postulate that the THC may be predominantly controlled by the climate forcing over the Southern Ocean on long glacial cycle timescales, but by the North Atlantic climate forcing on short millennial timescales (Liu, 2006). In terms of the more immediate concerns surrounding the anthropogenic forcing of climate triggering possible changes, recent ensemble model runs indicate a THC weakening of ~30% in response to freshwater input in the northern North Atlantic in the next 100 years (Stouffer *et al.*, 2006).

Particular concerns here are related to the recent dramatic warming in the Arctic and the implications of this for the Arctic Ocean heat budget in response to increased freshwater inputs. Some recent modelling work indicates Arctic Ocean freshwater export and volume are more sensitive to river runoff than sea-ice export (Rennermalm *et al.*, 2006). But no model, whether they be earth system models of intermediate complexity or fully coupled AOGCMs, simulates a complete shutdown (Dai *et al.*, 2005; Rennermalm *et al.*, 2006; Stouffer *et al.*, 2006).

1.2.2 The influence of the North Atlantic Oscillation (NAO)

The North Atlantic Oscillation (NAO) describes an atmospheric phenomenon in the North Atlantic sector as an organised motion of the Icelandic Low and Azores High (Defant, 1924; Walker and Bliss, 1932). These swings in the atmospheric sea level pressure difference between the Arctic and the sub-tropical Atlantic are most noticeable during the boreal cold season (October - March) and are associated with changes in the mean wind speed and direction. These changes alter the seasonal mean heat and moisture transport between the Atlantic and the neighbouring continents, as well as the intensity and number of storms, their paths, and their weather (Hurrell *et al.*, 2003) (Figure 1.4).

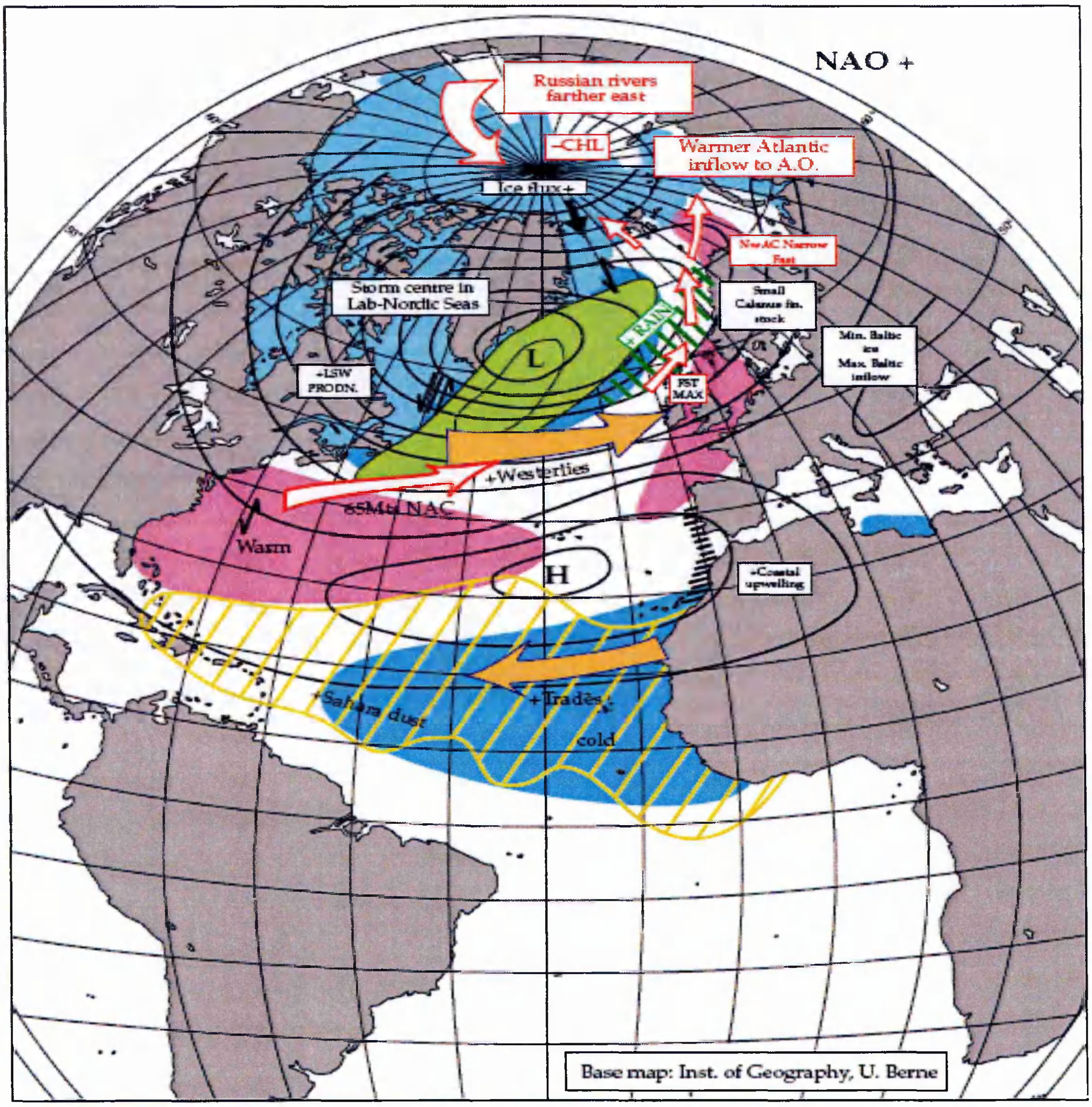


Figure 1.4: A schematic of the Atlantic-Arctic sector under NAO-positive conditions (Source: Hurrell and Dickson, 2004)

The NAO is one of the oldest known world weather patterns, with some of the earliest descriptions of it coming from seafaring Scandinavians several centuries ago (Hurrell *et al.*, 2003). In mid-latitudes the NAO is profoundly linked to the leading mode of variability of the whole Northern Hemisphere circulation, the annular mode or Arctic Oscillation (AO) (Marshall *et al.*, 2001; Cohen and Barlow, 2005). The AO may be thought of as the leading wintertime hemispheric low-frequency mode of variability in sea level pressure, while the NAO is the leading mode in the Atlantic basin (Marshall *et al.*, 2001).

Recent regional patterns of temperature change have been shown to be related, in part, to various phases of atmospheric-oceanic oscillations, such as the North Atlantic-Arctic Oscillation and possibly the Pacific Decadal Oscillation (IPCC, 2001; Cohen and Barlow, 2005). As a result, regional temperature trends (Section 3.4 presents original findings for the Highlands) over a few decades can be strongly influenced by regional variability in the climate system and can depart appreciably from a global average (IPCC, 2001).

Due to its influence on north-western European (and wider North Atlantic) climate, as well as marine and terrestrial ecosystems. Studies of the NAO have become central to the current global climate change debate, with research questions centred on to what extent anthropogenic climate change may alter the behaviour of the NAO, and hence the surrounding marine and land surface climate of adjacent regions such as the Highlands. The reader is directed to Sections 3.4.1 and 3.4.2 for further information on the NAO, as well as demonstrated linkages to regional temperature and precipitation variability as one of the original aspects of this work. The reader is also directed to Annex Paper 3 for further background information on NAO linkages to storm track variability, and the implications of possible future NAO-positive phase changes for the western coastal boundary of the Highlands.

1.3 Context and Background to The Altitudinal Modelling of Temperature

1.3.1 A global context

Mountain regions occupy about a quarter of the earth's surface and are home to approximately ten percent of the world's population (Becker and Bugmann, 2001: Reasoner *et al.*, 2001). Of considerable value as sources of food, fibre, minerals and water, they are also important repositories of carbon and soil nutrients as well as being rich in living natural resources, and thus comprise important refugia for biodiversity (Beniston, 2000, 2003; Reasoner *et al.*, 2001). Recognising the sensitivity of mountain environments and the consequences that changes in these environments might have, the scientific community has responded in recent years with a more focused interest in global change research in mountain regions (Becker and Bugmann, 2001; Reasoner *et al.*, 2001; Messerli, 2006), including the possible impacts of anthropogenic climate change.

From a scientific point of view, the strong altitudinal gradients in mountain regions provide unrivalled opportunities to detect and analyse global change processes, since:

- Associated with the compressed latitudinal life zones along the steep gradient stretching from the lowlands to the nival zone (Nagy *et al.*, 2003), meteorological (Section 5.2), hydrological, cryospheric and ecological conditions change strongly over relatively short distances. The boundaries between these systems are often climatically sensitive and thus have considerable value as indicators of, for example the impacts of wider climatic change (Becker and Bugmann, 2001; Reasoner *et al.*, 2001; Coll *et al.*, 2005).
- Related changes also impact upon the socio-economic infrastructure, including land-use and land-management practises, resource exploitation and the appeal of

mountain regions for tourism (Becker and Bugmann, 2001; Reasoner, *et al.*, 2001; Fredman and Heberlein, 2005).

The environmental problems already facing many mountain regions are only likely to be exacerbated by compounding issues such as anthropogenically induced changes to climate. If climate is viewed as the main ecological driving force (Gottfried *et al.*, 1999), there is consensus that montane ecosystems may be particularly sensitive to global warming (Hattenschwiler and Korner, 1995: Bunce *et al.,* 1998; Gordon, *et al.*, 1998; Gottfried *et al.*, 1999; Klanderud and Birks, 2003). The potential impact of warming on mountain environments is examined further in Chapters 5 and 6.

With many mountain plants intolerant to competition, fast-growing lowland species with broad altitudinal and ecological ranges are predicted to expand at the cost of slow-growing competition-intolerant species with narrow habitat demands (Holten, 1990, 1998; Korner, 1999, 2003; Klanderud and Birks, 2003). Given the limited extent of their current nival zone and hence limited scope for species migration upslope, maritime uplands are considered to be especially vulnerable to such climatically driven shifts. Figure 1.5 indicates the extent of Highland upland areas in a broader European mountain context.

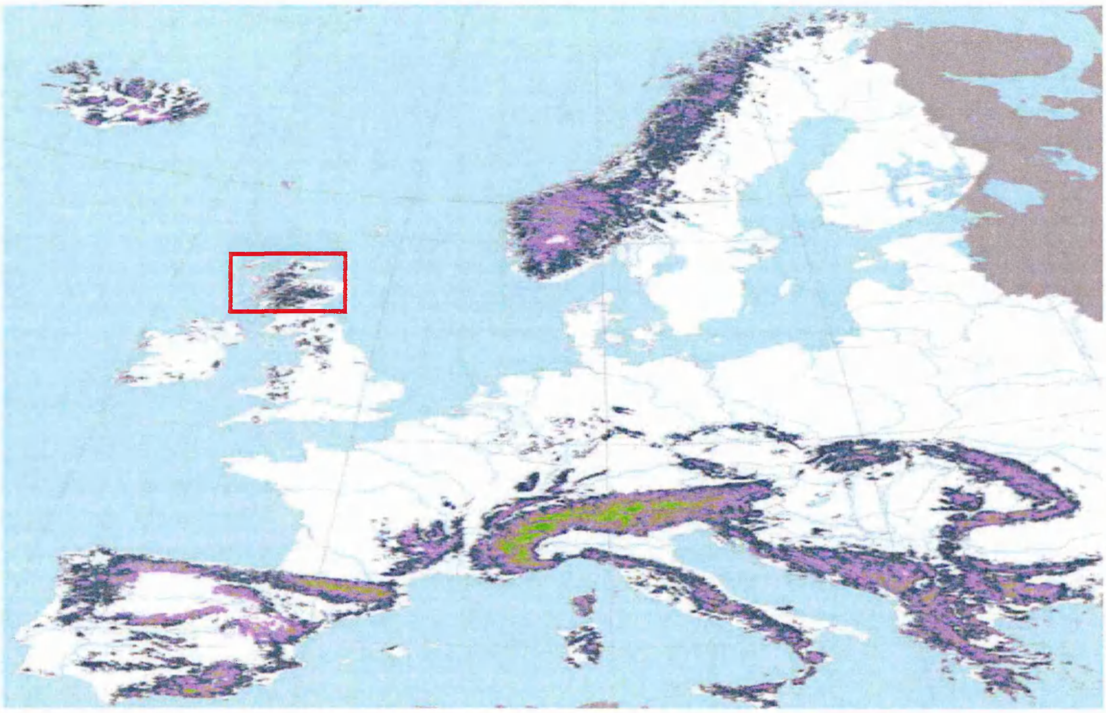


Figure 1.5: The Scottish Highlands (delineated in red) in a European upland context (Source: Kapos *et al.*, 2000).

1.3.2 Maritime upland sensitivity to climate change

Scottish maritime uplands are part of the boreal mountain zone which ranges from the uplands of Iceland through the Scottish Highlands, the Scandinavian mountains, the Urals, the higher northern part of the Central Siberian uplands and the mountain territories which occupy the south of Siberia and the Russian Far East. Birch woodlands (*Betula spp.)* dominate the European part of the zone, while Scots pine (*Pinus sylvestris)* is predominant in more continental eastern areas.

The woodlands of the Highlands may be regarded as forming an extreme western and highly maritime extension of the west European transition or ecotone from Temperate Deciduous (Summer) Forest, through Boreal Coniferous Forest, to Boreal Deciduous Forest. Above the tree line and associated with the altitudinal decrease in temperature and increasing effects of wind, these communities are replaced by boreal alpine as well as sub-nival and nival associations. In Iceland, sparse mountain pioneer vegetation occupies the highest altitudes while in the Highlands blanket bogs, heaths and dwarf shrub vegetation tend to be the dominants.

Situated on the seaward western edge of north-western Europe and subject to both maritime and continental influences, the climate of the Scottish Highlands is typified by spatial and temporal variability (Sections 3.1 and 3.2). Superimposed on these synoptic controls, orographic effects produce a locally variable climate across the region. With more than 4,500km^2 of the land surface being higher than 600m above sea level, such altitudinal gradients of change are an important control in the spatial pattern of climate across Scotland (Harrison and Kirkpatrick, 2001).

This variable climate contributes greatly to the biodiversity of the uplands, with a diverse mix of Atlantic, Arctic, Arctic-alpine and boreal elements occurring within a limited geographical area, and including many species on the edge of their global distributional range (Birks, 1997). Within this continuum of microclimates, most high altitude plant species are adapted to slow growth and are less competitive, with survival in the increasingly severe abiotic environment largely determined by the altitudinal range over which a species occurs.

Montane environments are among the least altered in Scotland and their arctic-alpine communities of plants and animals echo Scotland's post-glacial past, while moorlands, peatland and rough grassland form a mosaic of semi-natural habitats covering more than 50% of Scotland's land area (Jones *et al.*, 2002). Amid general concern that climatic changes will severely impact mountain regions, there is particular concern that the sensitivity of marginal maritime locations such as the Highlands may render them especially vulnerable to the impact of such changes (Pepin, 1995; Gordon *et al.*, 1998). Here the highest mountains are at the very edge of the alpine zone and their relatively mild, wet climate renders them particularly sensitive to climate change, particularly in the winter and spring half year. Any increase in oceanicity, with milder winters allied to varying and uncertain spring weather will create physiological and phenological complications for the survival of mountain plants.

While it might be expected that oceanic mountains would be buffered against climatic change by their more limited annual temperature range, by comparison with higher mountains such as the Alps, the nival zone is insufficient in extent to accommodate any potential upward migration of species (Crawford, 2000, 2001). With this reduced nival zone limiting the inherent climatic buffering capacity here by comparison with some of the higher European mountain regions.

Consequently under global warming scenarios, species with low migration potential are unlikely to achieve the relatively small distance migration which ensured their survival during preceding warmer periods of the Holocene. Winter climatic warming is a particular threat to montane floras in oceanic areas since the risks of premature bud burst and warm winter-induced carbohydrate consumption are greater (Crawford, 1997, 2000, 2001). Some of the strands above are explored more and in the context of original outputs from this work in Chapters 5 and 6.

1.3.3 Altitudinal zonation of vegetation in the Highlands

The natural altitudinal zonation of vegetation in the Scottish Highlands has been described by a number of authors (Smith 1900; Watt and Jones 1948; Poore and McVean 1957; Pearsall 1989; Thompson and Brown 1992; Brown *et al.* 1993; Gordon *et al.* 1998). Watt and Jones (1948), working in the Cairngorms, describe three main climatically limited vegetation zones, dominated by heather (*Calluna vulgaris*), crowberry (*Empetrium nigram*), blaeberry (*Vaccinium myrtillus*) and rush heath (*Juncus triffidus*) in order of increasing altitude. The main factors they considered to be driving the zonation of these plant communities were altitude and exposure, together with the extent and duration of snow cover.

Poore and McVean (1957) covered a wider area of the Highlands in their "new approach to Scottish mountain vegetation". In their view most mountain communities are interpreted as part of a climax mosaic, each being determined and maintained by powerful habitat factors. Nonetheless, they consider that in each community there is some cyclic change or micro-succession. However, the distribution and relative positions of communities is intrinsically linked to habitat factors (e.g. snow lie and exposure to wind). They describe a similar altitudinal zonation to Watt and Jones (1948), but also recognise the lowering of vegetation zones which occurs towards the north and west with increasing effects of oceanicity and moisture. Essentially they considered that most Scottish mountain communities fit into a framework in which the principal ecological controls are;

- Altitudinal Zonation; Oceanicity ; Snow Cover; Base Status; Moisture.

This early work shows that climate has a strong influence on vegetation zonation, with two important factors being temperature and winter wind speed through its influence on snow lie (Chapters 5 and 6). However, there is also a considerable latitudinal control and as a result, the montane (low to mid-alpine) zone lies above the potential tree-line of 700-800 metres in the more central parts of the eastern Highlands, but descends with latitude to 200 metres in the north and west (Thompson and Brown, 1992; Figure 2.4).

1.3.4 The influence of oceanicity on vegetation zones

Within the Highlands the degree of oceanicity increases westwards with maximum values along the western Atlantic seaboard (Crawford, 2000), while the east central Highlands are markedly more continental. Therefore, while one definition of the upland zone in Scotland is areas lying typically above the limits of enclosed farmland (Ratcliffe and Thompson, 1988), this definition can be extended to lower lying areas where climatic conditions are unfavourable, such as the exposed coasts of the north and north-west.

The pollen record demonstrates that this zonation has persisted for much of the Holocene, with Shetland, Orkney, the Outer Hebrides, parts of Caithness and some of the smaller Inner Hebridean islands having never supported extensive woodland even at low elevations probably as a consequence of exposure to sea spray, westerly gales and a short summer growing season (Birks, 1988). Whereas elsewhere across the region there have been more fluctuations in tree line in response to climate fluctuations throughout the Holocene. For example, during the early Holocene 9000 - 8000 radiocarbon years BP Mesocratic phase (*sensu* Birks, 1988) when temperatures were 2°C above present, the montane zone was much smaller and very local south of the Highlands (Thompson and Brown, 1992). Other studies have demonstrated that tree lines were significantly higher during the boreal period (5000 - 9000 radiocarbon years BP) and may have been up to 200 metres higher than present in the Cairngorms (Grace *et al*, 2002).

This pattern is coherent across northern Europe. In an analysis of tree-line chronology for *Pinus sylvestris* in the Swedish Scandes, Kullman and Kjallgren (2000) considered that treeline changes compared well with the Milankovitch model of orbital climate forcing, with an early Holocene thermal optimum followed by a period of cooler summers. Therefore, the expansion of the montane zone to its present extent was largely associated with the mid-Holocene Atlantic deterioration in climate and the associated increase in oceanicity.

The commonest meteorological assessment of oceanicity is mean annual temperature range. In some cases this can be used directly; in other cases it can be adjusted for latitude and related to defined extremes of oceanicity or its converse, continentality, as with Conrad's *Index of Continentality* (Conrad, 1946):

$$K = [1.7A/\sin(\phi + 10)] - 14$$

Where K is the index of continentality; A the average annual temperature range and ϕ, latitude.

In the British Isles this index gives a value of 12.5 for Heathrow, while the whole of Ireland and all but southern central and Highland Scotland can be described as oceanic, with values of < 8. The western coastal fringe of Scotland and the south and west of Ireland, together with the Northern Isles of Orkney and Shetland, can be described as 'hyper-oceanic', with values of mostly < 3 (Figure 1.6).

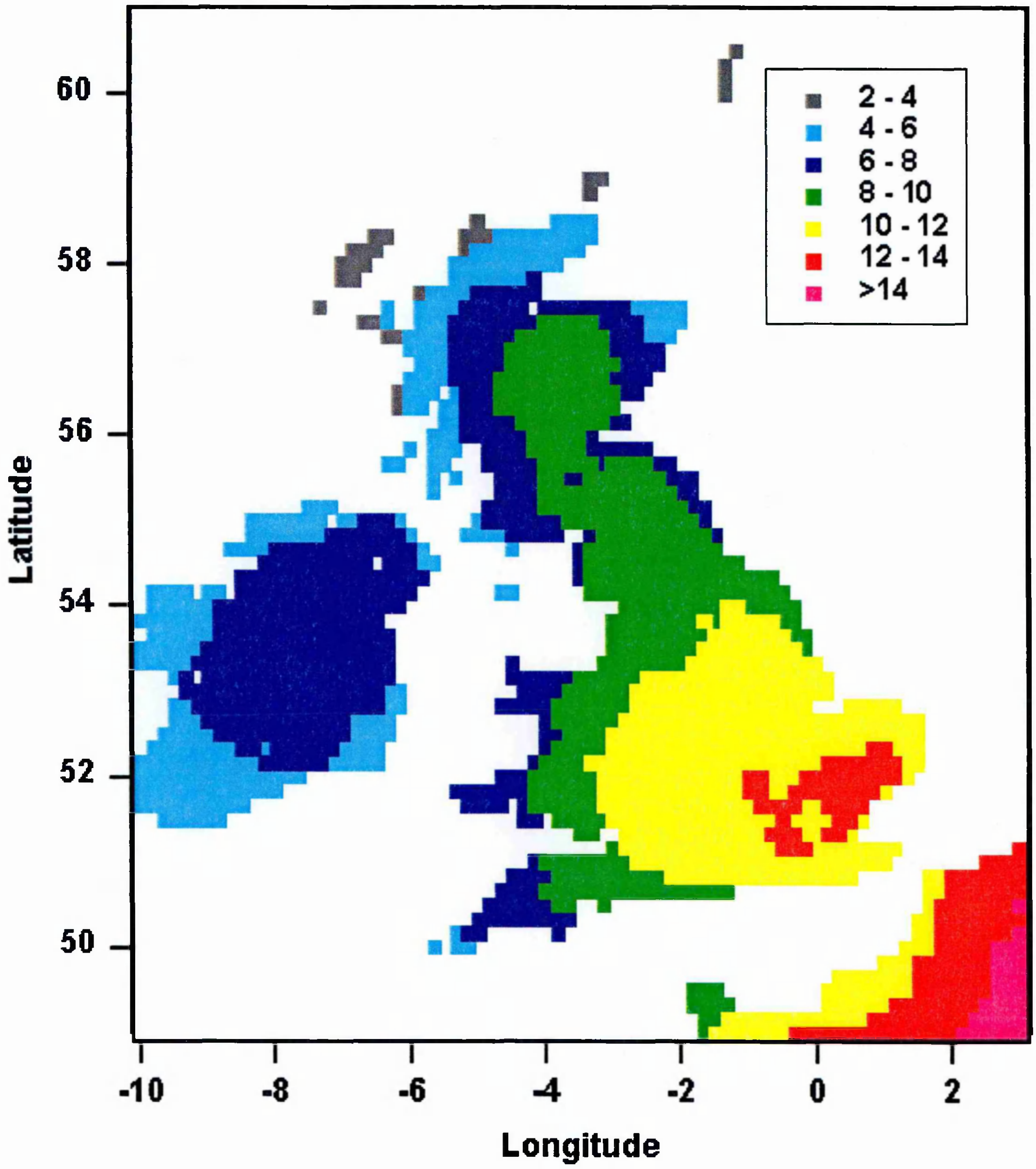


Figure 1.6: Conrad's index of continentality based on British Gridded Climatology for 1961-1990 (Crawford, 2002). Permission to use granted by RMM Crawford.

1.3.5 Species and habitat factors influencing migration in response to warming

Scottish mountains exhibit climatic variability both seasonally and spatially with the modifying factor of oceanicity strongly influenced by topography and aspect, as well as distance from the sea. Within this continuum of microclimates most high altitude plant species are adapted to slow growth and are less competitive, with the resulting trade-off between competitive ability and survival in the increasingly severe abiotic environment largely determining the altitudinal range over which a species occurs. In plant communities of generally slow growing species that occur in areas with severe abiotic conditions, competition may be limited, thus facilitation may play a more important role in community structure. If changes in climate lead to a reduction in the severity of the abiotic environment at high altitudes across the Highlands, this may in turn lead to an increase in competition among species and the possible invasion of species currently limited to lower altitude sites. An advancing tree line for example, or a denser forest below the tree line would have important implications for the global carbon cycle (increasing the terrestrial carbon sink) and for biodiversity of the alpine ecotone, possibly ousting rare species and disrupting alpine and arctic plant communities (Grace *et al.*, 2002).

Already tree lines in many mountainous regions have been responding to recent temperature changes (MacDonald *et al.* 1998; Kullman, 2000; Kullman and Kjallgren 2000; Marlow *et al.* 2000; Pallatt *et al.* 2000; Peterson and Peterson 2001; Klasner and Fagre 2002), with trees invading into meadows (Rochefort et al. 1994; Gavin and Brubaker, 1999; Wearne and Morgan 2001). Analogous changes in species richness and altitudinal distributions have occurred in Norwegian mountain plants in the last 70 years in response to relatively small changes in temperature (Klanderud and Birks, 2003; Kullman, 2003). While the high pine limit associated with the warm 20th century in the Swedish Scandes, stands out as an anomaly in the Holocene record (Kullman and Kjallgren, 2000; Kullman, 2001, 2002, 2003).

It has been estimated that overall, the predicted trend of mean temperature increase in the UK of about 1.5°C by 2050 is equivalent to a northward shift of 50-80 km per. decade or 40-55 m in altitude per decade (Gordon *et al.*, 1998). Similarly, it has been estimated that a rise in mean temperature of 1.8°C by 2100 would equate to an altitudinal descent of 300m for an area such as the Cairngorms (Hill *et al.*, 1999). However, results reported in subsequent sections of this work offer much improved estimates of possible future vegetation zone shifts in response to wider warming based on a more rigorous approach to modelling changes at a local scale (Chapter 5 and 6).

1.4 Uncertainty in Regional Climate Change Scenarios

1.4.1 GCM and scenario limitations

Overall, the IPCC place low confidence in the simulation of climate change at the regional scale ($10^5 - 10^7$ km^2) by available modelling tools due to;

- errors in the reproduction of present day regional climate characteristics;
- wide inter-model variability in the simulated climatic changes and the effects of important high resolution, sub-GCM grid-scale forcings and processes (Christensen, 2001).

Within the above scale a working definition of the regional scale could be described as the range $10^4 - 10^7$ km^2, with the upper end (10^7km^2) referred to as a sub-continental scale and scales smaller than 10^4km^2 being 'local' (Christensen, 2001).

While at global and continental scales Global Climate Models (GCMs) are able to reliably simulate important climatic mean features (Zorita and von Storch, 1999), their performance in reproducing regional climate is still characterised by systematic errors and limitations in accurately simulating regional climate conditions (Martin *et al.*, 1997; Mitchell and Hulme, 1999; Easterling *et al.*, 2000; Giorgi, 2005). Issues relating to the performance of climate models for a region like the Highlands are explored more fully in the Introduction to Chapter 4.

The reason for model-to-model differences in predictions lies largely in the construction of the models and how the modelling groups choose to parameterise different climate processes. Climate models attempt to represent the key elements of the climate system, e.g; atmosphere, ocean, land surface, cryosphere, and biogeochemical cycles. The largest part of climate change arises not from the direct effect of increasing GHGs but from the interaction between different components of the climate system, which give rise to a large number of positive and negative feedbacks (Jenkins and Lowe, 2003). Therefore climate model development carries with it the requirement for continuing evaluation of all aspects of the simulated climate (e.g the mean climatology, spatial and temporal variability, and extreme events), together with a thorough investigation of model systematic errors (Martin *et al.*, 2006; Ringer *et al.*, 2006).

As a result, regional climate prediction is a problem characterised by inherent uncertainty (Mitchell and Hulme, 1999; Hulme *et al.*, 2002; Jenkins and Lowe, 2003; Willows and Connell, 2003; Giorgi, 2005). Uncertainty in projected climate change can be attributed to three main sources;

- uncertainty in forcing scenarios;
- uncertainty in modelled responses to given forcing scenarios;
- and uncertainty due to missing or misrepresented physical climate processes in models (Hulme *et al.*, 2002; Jenkins and Lowe, 2003; Saelthun and Barkved, 2003).

In addition, recent work suggests that the largest uncertainty in future warming rates are over North America and Europe (Stott *et al.*, 2006).

However, assessments of the regional impacts of human-induced climate change on a wide range of social and environmental systems are fundamental for determining the appropriate regional policy responses to climate change (IPCC, 1998; Hulme *et al.*, 1999; Giorgi, 2005). Despite these requirements, it is generally accepted that the current generation of GCMs do not have the required spatial or temporal resolutions for direct assessments of future changes at the local scale (Gonzalez-Rouco *et al.*, 2000; Christensen, 2001; Schiermeier, 2004). Although recent developments have used a multi thousand member grand ensemble of simulations in a GCM to resolve regional change details (Stainforth *et al.*, 2005). With the uncertainty in the grand ensemble model response being explored as a step towards a probabilistic climate prediction system (Murphy *et al.*, 2004; Giorgi, 2005; Stainforth *et al.*, 2005; Tebaldi *et al.*, 2005).

Additional uncertainty in regional climate prediction stems from the inherent uncertainty surrounding regional and global climate system unpredictability and the complexity of processes and feedbacks that determine regional climate change (Mitchell and Hulme, 1999; Christensen, 2001; Jenkins and Lowe, 2003). While there is even greater uncertainty about changes in variability and extreme events (Smith *et al.*, 1998). Although recent advances have been made here using multiple model ensemble outputs (Christensen, 2002; Meehl and Tebaldi, 2004; Tebaldi *et al.*, 2005; Barnett *et al.*, 2006) and by inter-comparing Regional Climate Model (RCM) outputs (Christensen *et al.*, 2004; Deque *et al.*, 2005).

The limitations arising from imperfect scientific knowledge about the climate system are compounded by an inability to foresee how future GHG emissions will change. Future emissions from human activities depend upon socio-economic factors such as population, economic growth, and technology e.g. Since it cannot be known how these will change in the future, it is only possible to envision several plausible development trajectories over the next century (Table 2.1), and use models to estimate what emissions will be generated from these (Jenkins and Lowe, 2003). This has led to criticism from economists on the basis that the existing scenarios used for climate change projections rely on outdated economic theories which fail to reflect how lifestyle and energy demand in both rich and poor countries are likely to change (Schiermeier, 2006). These issues are briefly explored further in Chapter 8.

Adding to these problems, regional scale climate change impact assessments (CCIAs) are fraught with further difficulties such as;

- specifying the appropriate environmental response models and interpreting the impact results in the context of future socio-economic and technological change (Hulme *et al.*, 1999);
- since knowledge of impacts is largely based on historical experience and this knowledge is also imperfect, uncertainty about the socio-economic and environmental impact of future climate is compounded (Willows and Connell, 2003);
- a further complication is that many CCIAs implictly assume that each scenario represents just the signal of climate change and that multi-decadal natural climate variability can be ignored (Hulme *et al.*, 1999).

This is problematic since there is considerable variability (Section 3.4) in unforced thirty year climates and this natural climate variability can induce non-trivial responses in impact indicators. It thus becomes difficult to separate human-induced climate change impacts on natural resources from those attributable to multi-decadal natural climate variability, with the possibility that the estimated impacts will occur even in the absence of human-induced climate change (Hulme *et al.*, 1999).

Despite these sorts of issues the United Kingdom Climate Impacts Programme 2002 (UKCIP02) scenarios were produced using a state of the art double-nested RCM in a driving GCM. The modelling set up used to produce these and the characteristics of Regional Climate Models (RCMs) generally are reviewed more fully in the Introduction to Chapter 4.

1.4.2 Sources of uncertainty in the UKCIP02 scenarios

In UKCIP02 four scenarios were presented corresponding to four different possible pathways of future emissions, in order that emissions uncertainty could be taken into account (Hulme *et al.*, 2002; Jenkins and Lowe, 2003). The UKCIP02 scenarios can also be readily linked to a wider set of emissions and socio-economic scenarios (Table 2.1 and Section 2.4.1). Nonetheless, both science and model uncertainty remain, these in addition to the socio-economic and development uncertainties outlined above (and explored further in Chapter 8). The IPCC SRES scenarios corresponding to the four UKCIP02 scenarios (Table 2.1) are shown in Figure 1.7 below.

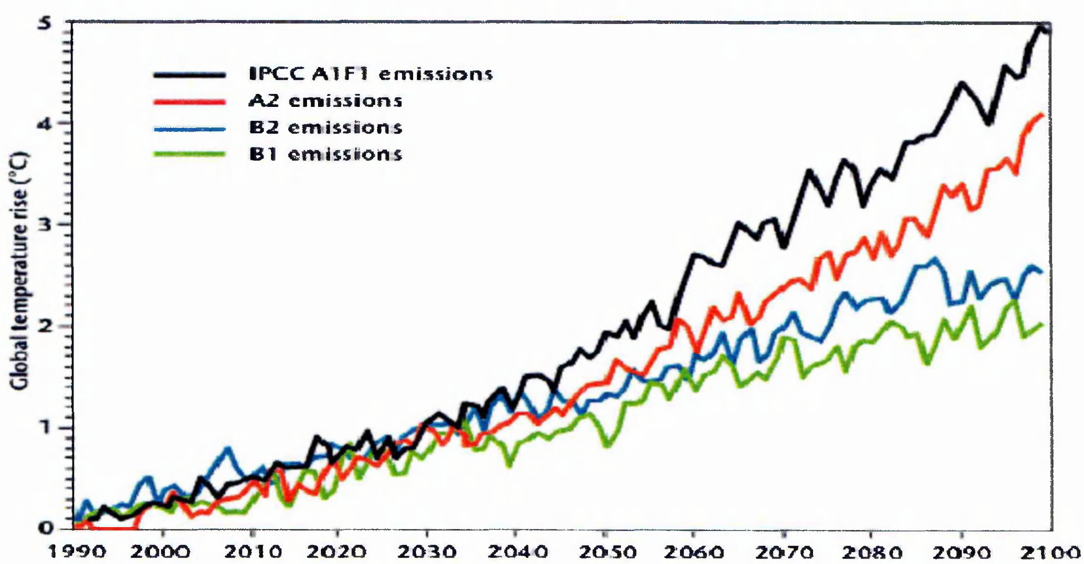


Figure 1.7: Global-mean surface-air temperature rise, estimated by the Hadley Centre HadCM3 model, resulting from the four emissions profiles (Source: Jenkins and Lowe, 2003).

It is apparent from Figure 1.7 that despite the immediate divergence of future emissions in the four scenarios, the warming over the next ~40 years from each emissions scenario is similar. This reflects the long effective lifetimes of CO_2 and the large inertia of the climate system locking warming in over the next few decades in response to current and historical emissions (Hulme *et al.*, 2002; Jenkins and Lowe, 2003). Whereas warming by the end of the century depends strongly on future emissions, with HadCM3 predicting a mean temperature change of +2°C - +5°C (Figure 1.7). Therefore emissions uncertainty makes little contribution to climate change uncertainty over the next forty or so years, whereas by the latter part of the century it is a major component of uncertainty (Jenkins and Lowe, 2003).

A further component of uncertainty is introduced from the dynamical downscaling process, this arising from the formulation of the RCMs themselves, HadRM3 in the case of the UKCIP02 scenarios. At the UK national scale this effect is small for seasonal mean air temperature, but more substantial for precipitation when nine RCMs driven by the same GCM are inter-compared (Rowell, 2004).

However, differences in predictions from different GCMs are generally much larger than the downscaling uncertainty associated with the RCMs. Therefore differences in results from the same RCM driven by different GCMs would be expected to be bigger than those from different RCMs driven by the same GCM, thus the largest uncertainty probably lies with the global prediction rather than the RCM downscaling (Jenkins and Lowe, 2003). The reader is also directed to the discussion section on the inter-comparison of RCM performance over a European domain in Section 7.5. Figure 1.8 below uses the case of winter and summer precipitation change over the UK to illustrate the substantial range of variation between nine AOCGCMs (see Notes below). Thus e.g.;

- winter change projections range from +1% for the PCM model to +61% in the NIES2;
- summer change projections range from –30% - + 4% (Jenkins and Lowe, 2003).

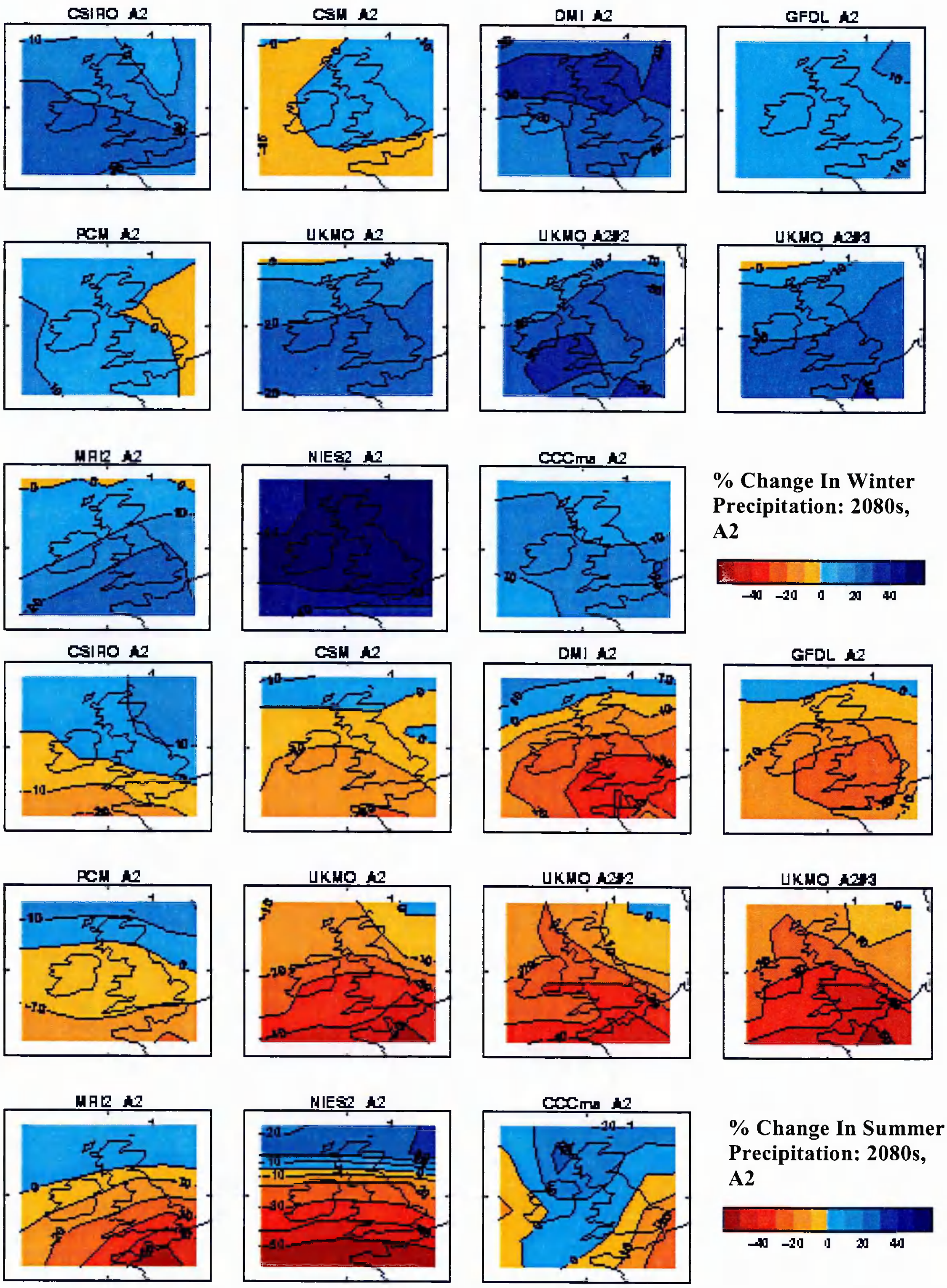


Figure 1.8: Change in winter and summer mean rainfall over the British Isles by the period 2071-2100 relative to 1961-1990 as predicted by nine climate models, all forced with the SRES A2 emissions scenarios (Table 2.1 and Section 2.4.1). There are three predictions from HadCM3 to illustrate natural variability (Source: Adapted from Jenkins and Lowe, 2003).

Model Acronym Notes, Figure 1.5

CSIRO: Commonwealth Scientific and Industrial Research Organisation (Australia)
CSM: National Centre for Atmospheric Research (USA)
DMI: Max Planck Institute, ECHAM4 (Germany)
GFDL: Geophysical Fluid Dynamics Laboratory (USA)
PCM: National Centre for Atmospheric Research (USA)
UKMO: Hadley Centre, HadCM3 (UK)
MR12: Meteorological Research Institute (Japan)
NIES2: National Institute for Environmental Studies (USA)
CCCma: Canadian Centre for Climate Modelling and Analysis (Canada)

1.5 Objectives

As has been reviewed here, the global climate is warming, with especially pronounced warming over landmasses and at high latitudes in the Northern Hemisphere. Despite ongoing debate as the climate change issue has become increasingly public and hence politicized, there is strong scientific consensus that the recent warming is largely anthropogenic in origin. While regional ‘surprises’ (e.g. a THC shutdown) remain a possibility, the consensus view is that recent warming trends will continue and amplify as the present century progresses in response to a continued build up of atmospheric GHG concentrations.

However, aside from the technical limitations associated with GCM and RCM outputs for a topographically diverse region such as the Highlands, there are numerous and wider underpinning uncertainties associated with projections of future climate change. This is especially true of obtaining locally valid projections of possible change for upland regions which are anthropocentrically perceived as vulnerable in terms of their future conservation and management. Consequently, in terms of policy development and contingency planning, there is a real need for developing methods which may deliver locally realistic projections of change for key climatic variables.

This thesis aims to address some of these difficulties and to arrive at a set of methods which allow possible climatic changes to be locally projected across a range of altitudes for the key climatic variables of temperature and precipitation. In pursuit of this goal a number of other region-specific difficulties in key areas will be tackled;

- Establishing baseline data of an appropriate record length and quality for a data sparse region of the UK that may be used to assess the performance of an RCM (Chapters 2, 3 and 4).

- Using the baseline data to illustrate aspects of sptial and temporal variation in Highland climate and relating these to important regional controls on climate variability, such as the NAO (Chapter 3).
- Developing and applying temperature lapse rate models (LRMs) to better assess RCM performance across the region, and a refinement of these for projecting future changes (Chapters 4 and 5).
- Exploring the potential impacts of outputs to broad upland vegetation zones, as well as in a case study context (Chapter 6).
- Considering wider issues in the pursuit of obtaining valid CCIAs in a local context. Arising from this some future challenges for the research community are identified (Chapters 7 and 8).

2. Methods

Construction Of A Baseline Climatology For The Highland Region

2.1 Creating A Temperature Station Network

2.1.1 Obtaining the raw data and determining a baseline period

Daily temperature data for mean maxima (T_{max}) and minima (T_{min}) were accessed for all UK Meteorological Office (Met. Office) weather stations via the British Atmospheric Data Centre (BADC) website (http://tornado.badc.rl.ac.uk); the BADC is the meta-data repository for the Met Office. The BADC database entries for the Highland region were identified and interrogated by county using the database 'WHEREIS' command to identify stations with an appropriate length of record for both T_{max} and T_{min}.

Stations were selected on the basis of having a largely intact daily or monthly long-term record spanning the 1961-1990 baseline period. In common with protocols in use elsewhere, a 30 year climatic average was required as this smoothes out many of the year to year variations in the climate. In addition, a 30 year average captures much of the inter-annual and short timescale variability of climate that may be relevant for an impact application and is of sufficient duration to encompass a range of weather anomalies (IPCC-TGCIA, 1999; Saelthun and Barkved, 2003).

This use of a 30 year average, as well as being consistent, is in accordance with the Intergovernmental Panel on Climate Change (IPCC) Task Group on Scenarios for Climate Impact Assessment (TGCIA) guidelines on using data for climate impact assessment (TGCIA, 1999). It has the further advantage of providing a common reference period for this study in accordance with the World Meteorological Organisation (WMO) defined 'normal' baseline (IPCC-TGCIA, 1999).

The 1961-1990 period generally has a fairly intact record of instrumental data and its use as a standard reference baseline period ensures there is consistency in climate change research work, both nationally and internationally (IPCC-TGCIA, 1999). It is also representative of the recent climate to which contemporary human or natural systems are likely to be reasonably well adapted (Saelthun and Barkved, 2003).

While the 1971-2000 period may have constituted a better baseline with the data for the 1990s capturing the warming and wetting trends of that decade, many of the station records for the Highland region available via the BADC were only updated to 1998/1999. In any case, most studies worldwide continue to use the 1961-1990 baseline owing to the greater availability of data compared to 1971-2000, and this period was used in the UKCIP02 scenarios to ensure consistency with previous UKCIP studies (Hulme *et al.*, 2002).

With the next official 30-year normal period adopted by the WMO set to be 1991-2020 thirty year record (Hulme *et. al.*, 2002), a combination of both protocol and data availability determined that the 1961-1990 baseline was the one to be used for purposes of this work. Also, following stakeholder consultation and feedback on what is required for the UKCIPNext scenarios (UKCIP Workshop, May 2006), it seems very likely that 1961-1990 will continue to be the reference baseline for the next set of Hadley Centre climate model outputs.

Another reason for avoiding a 1971-2000 baseline were some of the unusual climatic patterns recorded for the 1990s both globally, and at the UK-national and Scottish regional levels. For instance, it has been estimated that the mean temperature difference associated with using the 1971-2000 baseline (compared to 1961-1990) would be 0.3°C, at least for the south-east of the UK (Geoff Jenkins, pers comm.).

While as reported elsewhere in this work, a number of researchers have interpreted changes in temperature and precipitation over the 1990s as part of a pattern which may be part of a regional response to recent warming and an indicator of possible future changes in response to wider global warming. In the context of this interpretation, the use of a 1961-1990 baseline provides a more useful averaging period against which to assess trends through the 1990s. Taking a longer view, as regional instrumental records accrue through the early 21st century, trends in both temperature and precipitation can be scrutinised in relation to the 1961-1990 period.

Yearly data files for each station spanning the 1961-1990 baseline period were initially scrutinised prior to downloading. The size (kB) of intact files was also noted on the basis that files of a smaller size would likely be missing data; this was used to initially flag potential problem files prior to the further quality control procedures detailed below. Following these initial quality control procedures, data files for the baseline period were transferred to file directories organised by county and station name for further work.

2.1.2 Deriving the annual means

Station files were transferred to Excel worksheets in relevant folders and directories for ongoing work. Worksheets were kept discrete for each station year 1961-1990 in order to facilitate the next stage of the data quality control procedure. For each year and for all stations, default –999 entries were removed to avoid skewing the subsequently derived mean values. This procedure allied to the earlier file kB flag was used to identify station years which were incomplete or missing sizeable chunks of data. For each of the stations selected, the details of missing data and hence problem years were noted and used to construct Results Table 3.1.2 and summarised in a frequency plot (Figure 3.1.1).

As a standard procedure, years missing contiguous blocks of data extending beyond a week (as opposed to scattered missing values over the course of the year) were not used. The reasoning here was that artificial distortion to the annual mean as a result of missing values would also spuriously skew the long-term 1961-1990 baseline mean. Consequently, not all of the station long-term mean values are based on a full thirty year record for the 1961-1990 baseline. Results Table 3.1.1 provides a per station summary of the number of years used to construct the baseline mean values for each station. Following the above data quality control procedures, annual mean values for T_{max} and T_{min} were derived. These annual mean values were used to construct time series plots for each station in order to elucidate possible local variation in trends. Finally the annual mean values of T_{max} and T_{min} were summed and averaged to construct the baseline value for each station (Appendix Tables 3.1 and 3.2).

2.1.3 Deriving the seasonal means

For purposes of consistency with work undertaken elsewhere, climatological convention was followed in constructing the seasonal baseline mean values, thus;

- Winter = December, January, February
- Spring = March, April, May
- Summer = June, July, August

- Autumn = September, October, November

Again in accordance with convention, for winter, the values of the preceding December were used. Thus, for example, winter 1990 would comprise the December 1989 measurements together with the January and February values for 1990.

A Macro program (Visual Basic) was run for each of the station records spanning the baseline period to produce outputs for seasonal and annual means for T_{max} and T_{min}. The annual values obtained using semi-manual procedures were cross-referenced against program outputs in order to check and validate the performance of the program. Years known to be problematic in terms of missing data were manually removed from the program outputs. The program provided a useful flexibility in relation to this aspect of data quality control since it produced annual and seasonal mean values on an individual year basis for the 1961-1990 period. Consequently, when the 30 year means were compiled in the spreadsheets, previously identified problem years could be excluded.

Winter mean values were also spreadsheet compiled by adding the preceding December values (where month intact) to those of the January and February values for the subsequent year. The same spreadsheet-based procedures were used to obtain the winter 1961 mean values for those stations which had a data record for 1960.

An important decision was required in relation to the inclusion or otherwise of seasonal values for years in which the annual values could not safely be derived due to missing data. For instance, if only data corresponding to spring were missing and other seasons were intact, intact seasonal data for other part years could be used for deriving seasonal means. The decision largely centred on future utility of the station spreadsheet data and likely complications as it was read between software applications. As illustrated in Results Table 3.1.2, data gaps due to instrument or data logger failure had no consistent pattern, although the 1960s and 1970s appear to have been more problematic decades for some of the station records (Figure 3.1.1).

Consequently, the use of part-year data where the annual data was not used would result in considerable inhomogeneity of spreadsheet formatting, i.e. considerable inequality in column and row lengths. It was felt that this had considerable potential to create future difficulties and generate possible errors when reading data between applications. For example, some seasonal baseline means would be based on a longer yearly record than the <30 year annual means. Similarly, there would be between-season variation in the length of record used to calculate the baseline means.

Therefore, for purposes of consistency and ease of use, seasonal data were not used where the year had been rejected due to missing data. The only exception to this was the manual calculation of winter values described above, since this did not disrupt spreadsheet data structure or extend the yearly record used for this season relative to the baseline annual means.

2.1.4 Adding existing data to the station network

In order to extend spatial cover and increase the number of temperature stations used across the region, it was decided to obtain annual and seasonal thirty year means for the 1961-1990 baseline from existing Met Office summary data. Therefore, mean annual and seasonal T_{max} and T_{min} values were derived for the six stations:

1. Tiree; 2. Ardtalanaig; 3. Braemar; 4 Kinloss; 5. Stornoway; 6. Kirkwall.

It was reasoned that the addition of Tiree and Ardtalnaig would provide a better balance of observed temperatures to the south-west and south of the region when later assessing HadRM3's 1961 –1990 simulation against the observed station network. Similarly, it was reasoned that Stornoway and Kirkwall would provide a more balanced representation around the north-west and north-east for the same reasons, while the inclusion of Braemar and Kinloss increased the number of station records used to the east. Therefore data for the above stations were added to the station network and annual and seasonal mean values derived by manual procedures from the existing Met Office 1961-90 records.

2.2 Creating A Precipitation Station Network

2.2.1 Obtaining the raw data

Procedures here were largely the same as those for temperature detailed in Section 2.1.1 above.

2.2.2 Deriving the annual totals

Station files were transferred to Excel worksheets. Worksheets were kept discrete for each station year 1961-1990 in order to facilitate the next stage of the data quality control procedure. For total precipitation (Precip) data the control on BADC default (-999 values) was not required, monthly data was simply either present or absent. However, yearly data files were checked as for temperature in order to identify the kB flag on problem years. Years missing data for each station were recorded in Results Table 3.2.2. Again, following climatological convention, Precip totals for annual and seasonal values were used.

As with temperature, years in which data was missing were not used; in this instance, as data were from monthly stations, if a month was missing the year was rejected. The exception to this was use of December/January values (where these were available) from an otherwise non-intact year in order to derive winter seasonal mean values for adjacent years. Therefore, as for temperature not all station long-term mean values are based on a full thirty year record for the 1961-1990 baseline. Results Table 3.2.1 provides a summary of the number of years used to construct the baseline mean values by station. While Figure 3.2.1 summarises the frequency distribution of missing year records for all stations. Following these quality control procedures, annual totals were calculated and then summed and averaged to construct the baseline value for each station (Appendix Table 3.5).

2.2.3 Deriving the seasonal totals

As for temperature, standard climatological seasons were used, with calculation of winter totals following the same protocol. In order to derive spring, summer and autumn values, the raw data spreadsheets for the 1961-1990 period were opened simultaneously by station. This enabled a series of simple spreadsheet formulae to be embedded in order to derive the seasonal totals by year. These were then transferred to a spreadsheet summarising the baseline period mean seasonal values. This procedure was repeated for each of the fifty five stations and the ~thirty year means obtained. These are recorded in Table 3.2.1.

Obtaining winter values proved more problematic since the December values of the preceding year had to be used. In this case, the raw data spreadsheets covering the 1961-1990 period were opened simultaneously and the relevant values manually transferred to a pre-constructed grid. Data values were then re-entered to the summary spreadsheets and simple spreadsheet formulas used to derive the winter values. This procedure was repeated for each of the fifty five stations.

As for the temperature data, an important decision was required on whether or not to use partially complete years. As illustrated in Figure 3.2.1 and recorded in Table 3.2.2, with the exception of 1961 which appears to have been a problem year for stations across the region, data gaps due to instrument or data logger failure appear for various years. However, measurement/data recording problems appear to centre on the early-mid 1960s for a sizeable number of the stations (Figure 3.2.1).

Since monthly Precip totals are collected via a standard funnel type arrangement, it is assumed that some of the unusually snowy winters and springs associated with this period are the most likely source of the problem. If stations at different elevations were subject to drifting, this would result in mechanical obstruction of the funnel, while unusually low temperatures would also create collecting complications associated with freezing (see Section 3.2 for further discussion of this). It has also been generally noted that winter snowfall makes records from higher level gauges unreliable (McClatchey, 1996b).

As for temperature, considerations of consistency and homogeneity of spreadsheet structure determined that seasonal data were not used where the year had been rejected. The exception to this was some of the winter totals for certain stations where a lesser number of baseline years were used relative to other seasons. This was largely unavoidable and arose as a consequence of either December, January or February values being absent either side of an intact year. Again, this may have been a result of drifting problems affecting the collection arrangements as many of the gap years for these months are centred on the early to mid 1960s (Figure 3.2.1; Table 3.2.2).

2.3 Variability in Observed Temperature and Precipitation Data Time-Series

Data for individual stations were added together and averaged on a provisional intra-regional classification (Sections 3.4.1 and 3.4.2) to create approximate south-north station segments across the Highlands. These were used to examine variations in trends for selected values for west to east transects across the region for both temperature and precipitation (Section 3.4.1 and 3.4.2). Following the construction of time series plots for the regionally averaged data over the baseline period, a number of additional procedures were undertaken:

- Station annual T_{max} and T_{min} were summed and averaged with corrections being made to divisions to account for missing baseline years in the station records.
- Thirty year mean regional T_{max} and T_{min} for the 1961-1990 period were computed for the stations used in each sub-division. Similarly, annual regional means for the baseline period were computed for each year.
- Thirty year regional mean values for the years 1961-1990 were subtracted from the annual regional mean value. This allowing the difference to be expressed as an anomaly +/- with respect to the 1961-1990 regional mean in order to elucidate trends over the baseline period.

The procedure was applied to a suite of stations in a proximal west – east transect across the region and used to generate the selected anomaly plots in Section 3.4.1.

A similar approach was adopted for Precip stations. Data for individual stations were summed and averaged on a provisional intra-regional classification (Section 3.4.2) to create approximate south-north station segments across the Highlands. These were used to examine variations in trends for selected values for west to east transects across the region (Section 3.4.2). Derivation of anomaly plots relative to the 1961-

1990 baseline mean values largely followed the same procedure as for temperature. However, in this instance Precip totals were used to generate the selective anomaly plots presented in Results Section 3.4.2.

In order to place the 1961-1990 anomaly plots and fluctuations against a longer time-series for both temperature and Precip, some datasets were obtained from external sources;

- The annual land temperature records (1857-2003) are a Hadley Centre/Climatic Research Unit (CRU) dataset obtained via the Scottish Executive website and describe the differences from the 1961-1990 average for a 5 degree grid box extending from 55°-66° N and 0°-5° W (see Jones and Moberg, 2003). The annual precipitation records (1931-2003) were obtained from the same source and the methods are described in Jones and Conway (1997).

2.4 Working With The HadRM3 Model Outputs And Baseline Data

2.4.1 Obtaining the UKCIP02 data outputs from HadRM3

The full suite of HadRM3 variable outputs (Appendix Table 2.1) for the whole of the UK and the surrounding sea areas (Appendix Figure 2.1) were downloaded from the UKCIP02 website (ftp://[username]@tornado.eci.ox.ac.uk). Given the size of the files, WS_FTP Pro software was downloaded (http://www.ipswitch.com/) and used to facilitate the transfer. Data files for both the 50km x 50km gridded model outputs, together with the derived 5km x 5km grids were then read to appropriate directories. The UKCIP02 scenarios represent a considerable technical advance when compared to the UKCIP98 scenarios. In the UKCIP02 scenarios the UK land area is represented by 104 land grid cells at a 50km x 50km resolution and at varying elevations, compared to just 4 grid boxes in the 1998 scenarios (Hulme *et al.*, 2002).

The UKCIP02 scenarios can be linked to a wider set of emissions and socio-economic scenarios, Table 2.1 below provides a summary linking the UKCIP02 climate change scenarios to a number of these wider scenarios;

- in 1997 the IPCC set up a group to prepare a Special Report on Emissions Scenarios (SRES) based on four possible 'storylines' for future emissions;
- in 1999 the Foresight Programme of the UK Office of Science and Technology (OST) produced a further four storylines which were principally socio-economic in nature;
- while in 2001 UKCIP produced some UK specific socio-economic scenarios based on the Foresight work. This same generic scenario framework has been adopted by the Environment Agency (EA), although using a different nomenclature (Hulme *et al.*, 2002).

Table 2.1: Summary linking UKCIP02 climate change scenarios to other scenario developments (adapted from Hulme *et al.*, 2002)

SRES[1] Storyline	OST[2] Foresight Scenario	UKCIP[3] Socio-economic Scenario	Environment Agency Scenario	UKCIP02 Climate change Scenario
B1	Global Sustainability	Global Sustainability	Gamma	Low-Emissions
B2	Local Stewardship	Local Stewardship	Delta	Medium-Low Emissions
A2	Provincial Enterprise	National Enterprise	Alpha	Medium-High Emissions
A1F1	World Markets	World Markets	Beta	High Emissions

Notes:

[1] SRES: Special Report on Emissions Scenarios

[2] OST: UK Office of Science and Technology

2.4.2 Extracting the data for 'Highland' grid-boxes

The UKCIP02 HadRM3 index figure (Figure 2.1) was used to identify which model grid cells represented the Highlands for both 'land' and 'sea' areas immediately adjacent to the coast. Key attributes for 'sea' grid cells adjacent to the coast for the Highlands are summarised in Appendix Table 2.2. While attributes for the 'land' grid cells corresponding to the Highlands are summarised in Table 2.2 below; latitude and longitude co-ordinates reference the centre point of the HadRM3 50km x 50km grid cells (Hulme *et al.*, 2002).

Table 2.2: Spatial Reference and Elevation Information For HadRM3 Highland 'Land' Boxes

HadRM3 Grid Box Number	Latitude(°N)	Longitude (°W)	Elevation (m)
92	58.55	-4.48	100.35
93	58.63	-3.66	48.35
108	58.03	-5.13	252.02
109	58.11	-4.32	202.93
110	58.20	-3.50	68.58
124	57.50	-5.75	277.78
125	57.60	-4.96	384.65
126	57.68	-4.16	151.23
142	57.07	-5.58	445.07
143	57.17	-4.79	438.53
144	57.25	-4.00	412.19
145	57.33	-3.20	320.73
146	57.41	-2.40	113.38
160	56.64	-5.41	460.54
161	56.73	-4.63	618.85
162	56.82	-3.84	479.38
163	56.90	-3.06	339.39
178	56.21	-5.24	333.26
179	56.30	-4.47	365.22
180	56.39	-3.69	201.68
181	56.47	-2.91	77.14

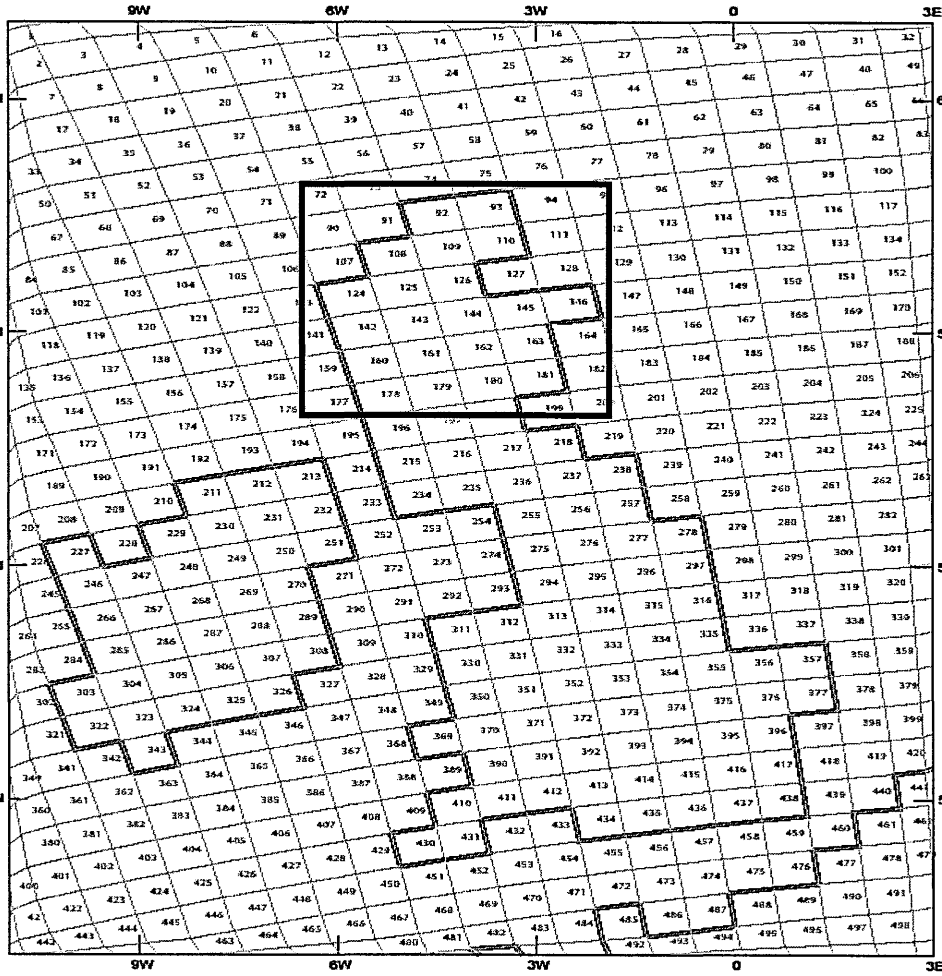


Figure 2.1: UK 50km x 50km land and sea grid cells - HadRM3 (Source: Hulme *et al.*, 2002). 'Highland' land grid cells are annotated.

The HadRM3 data for all the UK grid boxes was read to Excel and the data for Highland grid boxes transferred to a separate series of spreadsheets. HadRM3 data for both T_{max}, T_{min} and Precip for each of the four scenarios (Low, Medium-Low, Medium-High and High) and three future time slices (2020s, 2050s and 2080s) were extracted and read to discrete directories and folders to create a regional database for these variables. Similarly, HadRM3 data outputs for the same variables based on the model's simulation of the 1961-1990 baseline for the Highland grid cells was extracted and read to a directory.

What the UKCIP02 scenarios and future time slices provide are;

- temperature change ($\Delta T^{o}C$) relative to the 1961-1990 baseline simulation at a monthly, seasonal and annual resolution;
- and precipitation change ($\Delta\%$) at the same temporal resolution relative to the same baseline period.

A number of additional computational procedures were undertaken to link the data for future scenarios and time slices for the Highland grid cells (Table 2.2) to those of the outputs for the simulated 1961-1990 baseline. These were required in order to convert the Δ future change for T_{max}, T_{min} and Precip to a whole annual and seasonal value relative to 1961-1990 for purposes of subsequent work.

2.4.3 Spatially referencing stations to the HadRM3 grid cells

The latitude and longitude co-ordinates for the centre point of each HadRM3 grid cell corresponding to the Highlands were input to a spreadsheet matrix with longitude and latitude axes defined at a resolution of 0.1°. Longitude was converted to a distance equivalent (km) in order to define the matrix area covered by each HadRM3 grid using the relationship:

(1) 1.0° Longitude = 111.32km, therefore 0.1° Longitude = 11.13km

Based on the above conversion, each HadRM3 50km x 50km grid cell is equivalent to 50/11.13 = 4.49 spreadsheet cells or 2.25 cells in each direction from the centre point co-ordinates in the matrix. The above procedure was repeated for latitude using the relationships;

(2) 1.0° Latitude = 110.57 km at equator; 1.0° Latitude = 111.70 km at poles

Therefore the difference of 1.13km above equates to a proportional range difference of ~0.63km - ~0.67km at latitudes 56°N – 59°N.

Following the above calculations, values were converted *pro-rata* to an equivalent 0.1° distance resolution and corrected systematically for latitudes 56°N - 59°N to allow for the slight flattening towards the pole increasing the degree length. Given the relatively coarse (0.1°) resolution of the matrix, the above corrections for latitude still equated to ~2.25 spreadsheet cells in each direction from the HadRM3 grid cell centre point co-ordinates. Temperature and precipitation stations used to construct the observed baseline were then input by latitude and longitude co-ordinates to the same matrix and a simple nearest neighbour method used to interpolate stations to the nearest HadRM3 grid cell. The approach used here is illustrated in Appendice Figure 2.1.

2.4.4 Inter-comparison: Station Observed 1961-90 Baseline Averages And The HadRM3 Simulated 1961-90 Averages

2.4.4.1 Temperature: Method I Inter-comparison

In this analysis, all nineteen temperature station records for the 1961-1990 baseline period were included in the analysis, including the two island sites at Isle of Rhum and at Prabost on Skye. In this analysis, no attempt was made to infill missing years for incomplete station records, this in order to compare differences when infilled values were used (Section 3.3.1). Consequently the actual record of years used in the baseline period ranged from 23-30 years. Table 3.1.1 details the number of years used for each station.

For the reasons discussed in Section 2.1.4, existing Met Office station data were added for six stations, including three mainland stations (Ardtalnaig, Braemar and Kinloss) island stations of Tiree, Stornoway and Kirkwall providing a total of twenty five observed station records for comparison with the corresponding HadRM3 grid simulated 1961-1990 values (Figure 2.2). For purposes of undertaking the inter-comparison, island station sites were referenced to the closest HadRM3 'mainland' grid-cell via the spreadsheet matrix.

In this analysis no attempt was made to use lapse-rate values to adjust 1961-1990 observed station values to the elevation of the corresponding HadRM3 grid-cell. HadRM3 grid cell-simulated 1961-1990 values were simply subtracted from the corresponding observed station values for selected annual and seasonal maxima and minima, with the differences plotted by longitude and season (Section 4.1.1; Figure 4.1).

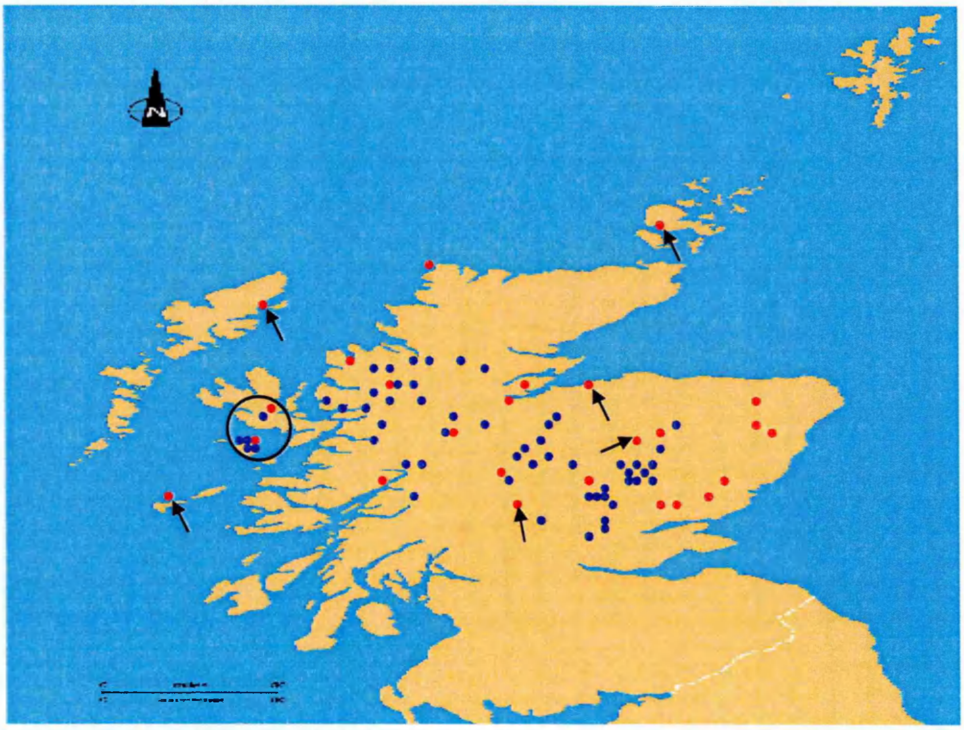


Figure 2.2: Outline map indicating station locations. Temperature stations are denoted in red, precipitation stations in blue. Arrows indicate Met Office station records added, while island sites dropped from subsequent analysis are circled.

2.4.4.2 Temperature: Method II Inter-comparison

For this analysis, and to counter problem of gaps in the observed 1961-1990 baseline record, annual values of T_{max} and T_{min} for the nineteen station records were cross-correlated in a MINITAB correlation matrix. Those stations with data sets exhibiting the highest Pearson Correlation Coefficients and lowest p-values (ranges below) were identified and used to infill gap years for stations with an incomplete 1961-1990 record:

- T_{max}, $r = 0.725\text{-}0.960$, $p < 0.001$
- T_{min}, $r = 0.533\text{-}0.930$, $p < 0.001$ - < 0.006

Coefficient values and p-values for individual stations, together with information on the station records used for infilling are recorded in Table 3.3.1.

Following these procedures, island station values were dropped from the analysis on the basis that island sites were not covered by the UKCIP02 HadRM3 grid cells and their inclusion therefore biased the inter-comparison against HadRM3 grid 1961-1990 simulations. For similar reasons the data for the existing Met Office island station records (Tiree, Stornoway and Kirkwall) were dropped from the analysis and only the three mainland station data sets (Ardtalnaig, Braemar and Kinloss) retained.

As an additional measure, and in recognition of the considerable disparity between the elevation of the HadRM3 grids compared to the altitude of the stations, some form of altitudinal correction had to be undertaken if the observed and HadRM3 simulated baselines were to be inter-compared with any validity. Therefore, to correct the mismatch in elevation, annual and seasonal temperature lapse-rate values were used to relate station altitudes (nearest 10 metres) to the corresponding HadRM3 grid for annual and seasonal values. In this approach only mean annual and seasonal lapse-rate values of the values in Table 2.3 were applied. As for the previous experiment, HadRM3 grid outputs for the 1961-1990 baseline were subtracted from the corresponding observed station values and the results plotted for seasonal values (Section 4.1.2, Figure 4.3).

Table 2.3 Annual and seasonal range of temperature lapse rate values. (Source: Dr S.J. Harrison, based on previous work undertaken in the Scottish uplands).

Season	**Lapse Rate °C/1,000 m**	
	Maximum Temperature (T_{max})	**Minimum Temperature (T_{min})**
Annual	9.5-10.5	4.0-7.0
Spring	9.8-10.2	6.0–8.0
Summer	9.5-9.8	5.0–7.0
Autumn	9.2-10.0	3.0–5.0
Winter	9.5-10.2	3.0–8.0

2.4.4.3 Precipitation: Method I Inter-comparison

All fifty five monthly precipitation station records for the 1961-1990 baseline period were included in this analysis, including the sites on Rum and the one on Skye. No attempt was made to infill missing years for incomplete station records, therefore the actual record of years used to construct the observed baseline ranged from 21-29 years; the number of years used for each station are detailed in Table 3.2.1.

As per the methods described for temperature, island station sites were referenced to the closest 'mainland' HadRM3 grid-cell via the spreadsheet matrix. HadRM3 grid cell-simulated 1961-1990 values then were simply subtracted from the corresponding observed station values for selected annual and seasonal Precip totals and the results plotted (Section 4.2.3, Figure 4.5).

2.4.4.3 Precipitation: Method II Inter-comparison

In this analysis MINITAB was again used to cross-correlate annual values of precipitation for all fifty five stations in a correlation matrix. As for temperature, stations with data sets exhibiting the highest Correlation Coefficients (Pearson) and lowest p-values were identified and used to infill gap years in the 1961-1990 record ($r = 0.725 - 0.983$, $p < 0.001 - < 0.005$).

Despite these measures, monthly rain gauge records for 1961 were largely unobtainable (see explanation offered in Sections 2.2.3 & 3.2); consequently, the precipitation baseline could only be constructed for the 29 year baseline period of 1962-1990. Having applied these procedures, island stations were excluded from the inter-comparison with the HadRM3 grid-simulated 1961-1990 mean values on the basis that their exclusion would allow a more valid inter-comparison of values. As for previous experiments, HadRM3 grid outputs were subtracted from the corresponding observed station values and the results plotted for selected annual and seasonal values (Section 4.2.4, Figure 4.6).

2.4.4.4 Precipitation: Method III Inter-comparison

Altitudinal adjustment was applied to selected station data in order to adjust observed mean values for some component of orographic enhancement. While recognising that orographic effects are highly localised in the Highlands and that air-mass characteristics are an important control (M^{c}Latchey, 1996; Brunsdon *et al.*, 2001), nonetheless some form of adjustment of station mean values to an altitude corresponding to HadRM3 grid elevations was desirable in order to more realistically inter-compare values.

Consequently, rates of rainfall increase with elevation deemed 'typical' for the region were obtained (Weston and Roy, 1994) and used to adjust selected station values (nearest metre) to the same elevation as the corresponding HadRM3 grid cell. For the subsequent inter-comparison analysis, a number of refinements were made to this approach:

- Recognising that validating the performance of climate models requires an intensive network of observational data, following altitudinal adjustments to account for orography, station data were averaged by area under the corresponding HadRM3 grid cell.
- In order to keep spreadsheet data arrays manageable, the above procedures were applied selectively for three HadRM3 grid cells (Had Grid IDs 125, 144, 163: Figure 2.1) on a west to east transect across the region.

Selection of the HadRM3 grids for evaluation using this approach were largely determined by the spatial distribution of the available observed station records. Working on the basis that more observed 1961-1990 data beneath the corresponding HadRM3 grid cells would provide a better baseline for purposes of inter-comparison, the spreadsheet matrix was re-visited and two selection criteria based on availability of observed station records were applied to the choice of HadRM3 grids:

(i) The greatest number of available station records corresponding to the HadRM3 grid dimensions.

(ii) The proximity of mean station elevations to that of the corresponding HadRM3 grid, this to keep the level of adjustment for orography (with accompanying assumptions) to a minimum and hence improve the validity of the inter-comparison.

For consistency with previous experiments, HadRM3 grid outputs were subtracted from the areally averaged and altitudinally adjusted station values and results plotted for mean seasonal values (Section 4.2.5, Figure 4.7).

2.5 Lapse Rate Models (LRMs) for The Projection of Possible Future Temperature Changes

2.5.1 Obtaining upland station temperature records for validating the lapse rate models

There is not an extensive data repository of upland climate data for the Highlands and there is certainly no station with an intact record for the 1961-1990 baseline period. Nonetheless, a search of the BADC database revealed four possible candidate stations with records of variable lengths (record details are in Table 5.1):

1. Cairngorm, 663 metres. 2. Cairnwell, 933 metres. 3. Aonach Mor, 1033 metres.
4. Cairngorm Siesaws, 1245 metres.

Following similar download and file organisation protocols applied when obtaining the baseline data, initial data quality control procedures adopted in preceding Sections 2.1.2 and 2.1.3 were replicated for each of the station records. Following from these procedures the same Macro program was run for each station record to obtain annual and seasonal mean T_{max} and T_{min} for each year in the record. These were then summed and averaged to produce mean annual and seasonal maxima and minima across the length of the existing record (Results Table 5.2).

2.5.2 Constructing and testing the lapse rate models against the 1960-1990 baselines (observed and HadRM3 simulated)

A range of annual and seasonal lapse rate values representative of mean annual and seasonal T_{max} and T_{min} were converted to 10 metre and 50 metre increments (Table 2.4 and Table 2.5). Following these adjustments, spreadsheet models were constructed at a 50 metre increment across the annual and seasonal range of shallow and steep temperature lapse rate values for both T_{max} and T_{min}.

Table 2.4: T_{max} values, 10 metre and 50 metre increments for shallow and steep temperature lapse rate values applied

	10 m Shallow Rate (°C)	10 m Steep Rate (°C)	50 m Shallow Rate (°C)	50 m Steep Rate (°C)
Annual	0.095	0.105	0.480	0.530
Spring	0.098	0.102	0.490	0.510
Summer	0.095	0.098	0.480	0.490
Autumn	0.092	0.100	0.460	0.500
Winter	0.095	0.102	0.480	0.510

Table 2.5: T_{min} Values, 10 metre and 50 metre increments for shallow and steep temperature lapse rate values applied

	10m Shallow Rate (°C)	10m Steep Rate (°C)	50m Shallow Rate (°C)	50m Steep Rate (°C)
Annual	0.040	0.070	0.200	0.350
Spring	0.060	0.080	0.300	0.400
Summer	0.050	0.070	0.250	0.350
Autumn	0.030	0.050	0.150	0.250
Winter	0.030	0.080	0.150	0.400

Following these procedures, observed station values with data quality controlled for the 1961-1990 baseline were obtained for representative western and eastern stations (Onich and Balmoral respectively). Data outputs for HadRM3's simulation of mean annual and seasonal T_{max} and T_{min} for the 1961-1990 baseline for the corresponding HadRM3 grids were obtained from the UKCIP02 database. It should be noted that there are altitudinal differences between these stations and the corresponding HadRM3 grids;

- Onich = 15 metres: HadRM3 Grid 160 = 460.54 metres. Altitudinal difference = 445.54 metres.

- Balmoral = 283 metres: HadRM3 Grid 145 = 320.73 metres. Altitudinal difference = 37.73 metres.

These differences necessitated a number of additional computing procedures prior to applying the lapse rate models to both the station observed and HadRM3 simulated 1961-1990 baseline values.

For the HadRM3 grid 160 annual and seasonal mean output values, the 10 metre increment observed lapse rate values were used to adjust the grid cell elevation back to 450 metre equivalents, this to facilitate the convenient application of the 10m and 50m lapse rate values in the models. While for the Onich station values, the 10 metre increment lapse rate values were used to adjust annual and seasonal mean values to an equivalent elevation of 50 metres. Again this was for purposes of utility in running the lapse rate models using the 50 m incremental values. Following this adjustment the 50 metre increment values were used to deductively lapse adjust the annual and seasonal Onich values for T_{max} and T_{min} to 1300 metres (Results Section 5.2.1). In all of these experiments, model adjustments were made to a common elevation of 1300 metres since this approximated to the highest hills west and east; Ben Nevis, 1345 m and Beinn Macdui, 1309 m respectively.

This procedure was repeated for the HadRM3 grid 160 simulated values, with the 50 metre increment models being used deductively to lapse adjust values to 1300 metres from the 450 metre grid adjusted values. While the models were again used deductively to adjust the HadRM3 grid outputs back to a 50 metre equivalent, this to allow comparison across the full altitudinal range with the Onich lapse adjusted station values (Results Section 5.2.1).

For the Balmoral station values, the 10 metre increment lapse rate values were used deductively to adjust annual and seasonal mean values to an equivalent elevation of 300 metres. While for the HadRM3 grid 145 annual and seasonal value outputs, the 10 metre increment lapse rate values were used additively to adjust these to an equivalent elevation of 300 metres. Following from these adjustments, the 50 metre increment models were used deductively to lapse adjust and project both the HadRM3 simulated and station observed values to 1300 metres (Results Section 5.2.2). Aspects of procedures described here are represented schematically in Figure 2.3 below.

2.5.3 Perturbing the baseline model outputs with selected future scenario outputs from HadRM3

Prior to selecting the initial HadRM3 simulated future temperature changes relative to 1961-1990 ($\Delta T^{o}C$) changes for use in perturbing the baseline lapse rate models, a number of scenario selection criteria were applied. It was considered that these should represent a number of future time slices, as well as representing a range of possible future socio-economic trajectories and hence differing levels of uncertainty. Consequently, three initial scenarios were selected ranging from the 2020s to the 2050s and through to the 2080s. These are linked to their equivalent IPCC SRES storylines and OST Foresight Scenarios (Table 2.1) as below;

- 2020s Low: ~SRES BI Scenario and OST Foresight Scenario = 'Global Sustainability';
- 2050s Medium-Low: ~SRES B2 Scenario and OST Foresight Scenario = 'Local Stewardship';
- 2080s High: ~SRES A1F1 Scenario and OFT Foresight Scenario = 'World Markets'.

It was considered that the selection of these scenarios spanned the range of possible future development trajectories and could, therefore, be considered representative of a wide range of possible futures on different time-scales. However, in results Sections 5.3 and 5.4 experimental outputs from these aspects of the modelling focus on the 2050s and 2080s since the magnitude of prospective future changes in temperature with altitude are greater, and hence potentially more significant in terms of their impact.

Schematic Figure 2.3 below is intended to help the reader interpret the description of methods to follow here, as well as those described in Section 2.5.4. The 10 metre increment lapse rate values were used deductively to adjust the Onich 1961 – 1990 annual and seasonal mean values to 50 metres across the range of lapse rate values described in Table 2.4. Thereafter, the 50 metre values were applied deductively to project the observed values to 1300 metres (for reasons discussed above) across the range of annual and seasonal lapse rate values in Tables 2.4 and 2.5.

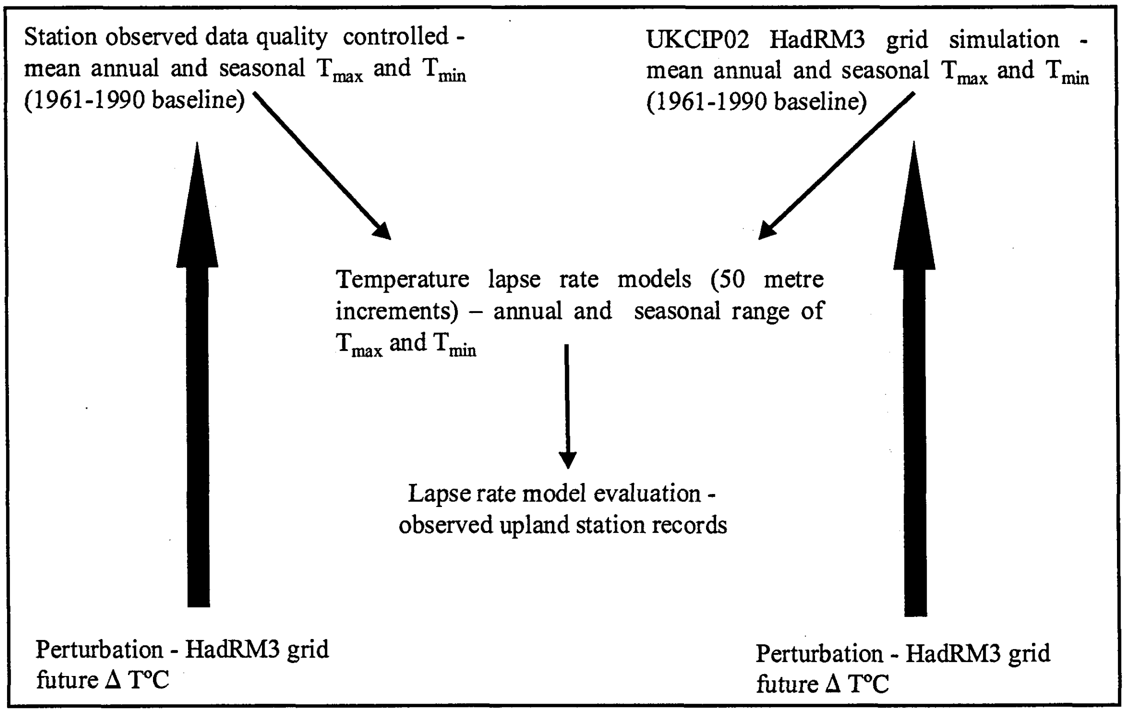


Figure 2.3: Schematic of method used to validate and construct lapse-rate models. Thin arrows indicate data handling procedures for observed and HadRM3 simulated 1961-1990 baseline data. Thick arrows schematically indicate lapse rate model perturbation using HadRM3 future scenario outputs.

For the HadRM3 Grid Cell 160 simulated 1961 -1990 values, the 10 metre lapse rate values were used additively to adjust outputs to a 450 metre equivalent; thereafter the 50 metre lapse rate values were used deductively to lapse adjust the baseline values to a 1300 metre equivalent elevation. It should be noted that for the experimental outputs presented in Sections 5.3.2 and 5.3.3, the projection of HadRM3 1961 – 1990 values to < 460 metres were derived by additively applying the 50 metre lapse rate increment models to adjust the HadRM3 grid outputs back to a common elevation of 50 metres.

Following these procedures and in order to project ΔT°C perturbed Onich 1961 -1990 baseline values across the elevational range to 1300m, Had Grid 160 ΔT°C outputs for both the 2050s and 2080s were additively adjusted to 50 metres using a combination of the 50 metre and 10 metre lapse rate values. After these adjustments, the 1961 – 1990 baseline values were additively perturbed by ΔT°C values from the HadRM3 grid for both scenarios and the 50 metre increment lapse rate models used to project future changes to 1300 metres (Sections 5.3 and 5.4).

By contrast, when applying the ΔT°C values for both scenarios to the HadRM3 1961 – 1990 grid-simulated values, the 10 metre lapse rate values were used to perform a simple additive adjustment (to 450 metres). Thereafter the 50 metre increment models were applied deductively from 450 metres to run the future ΔT°C perturbed values presented in Sections 5.3 and 5.4. Consequently, it should be noted that experimental results presented in Sections 5.3 and 5.4 projecting perturbed HadRM3 grid-simulated values back to 50 metres were derived by additively applying the 50 metre increment lapse rate values across the ranges described in Tables 2.4 and 2.5.

For the projection of possible future ΔT°C changes in the eastern uplands, a very similar set of procedures were employed. In this case, the 10 metre lapse rate values were used to adjust the Balmoral observed and HadRM3 Grid 145 simulated 1961 - 1990 values to a common elevation of 300 metres. Following these adjustments, the 50 metre increment lapse rate models were applies deductively to project both the observed and simulated 1961 – 1990 baseline values to 1300 metres (Sections 5.3 and 5.4).

In order to altitudinally project HadRM3 ΔT°C perturbed values for the Balmoral baseline, outputs of ΔT°C from HadRM3 Grid 145 for both he 2050s and 2080s were additively adjusted to a 300 metre equivalent elevation using the 10 metre lapse rate values. Following these adjustments, the 50 metre increment lapse rate models were used deductively to project the $\Delta T^{o}C$ perturbed values to 1300 metres (Sections 5.3 and 5.4). Similarly, given that the HadRM3 Grid 145 simulated outputs for 1961 - 1990 had already been additively corrected to 300 metres. The $\Delta T^{o}C$ outputs for both future scenarios were additively adjusted to a 300 metre equivalent value and the 50 metre increment lapse rate models used deductively to project the $\Delta T^{o}C$ perturbed values (Sections 5.3 and 5.4).

2.5.4 Utilising the lapse rate models to infer future shifts in key seasonal isotherms

In contrast to the methods employed in Section 2.5.2 where models were constructed to project across the range of annual and seasonal values associated with shallow and steep lapse rate values. For this experiment, only mean annual and seasonal lapse rate values were used and 50 metre increment spreadsheet models again constructed to a cut-off point of 1300 metres to more or less coincide with the elevation of the highest hills east and west. Following model construction, the 1961 - 1990 mean annual and seasonal values for Onich in the west and Balmoral in the east were adjusted using the lapse rate values to an elevation of 1300 metres.

Approximate upper limits associated with vegetation zones for both western and eastern mountains were identified (Figure 2.4). For purposes of this experiment, the upper elevation limit associated with each zone were taken to be;

- 300 metres in the west and 600 metres in the east for the forest zone; 600 metres in the west and 900 metres in the east for the sub-alpine zone; 900 metres in the west and 1200 metres in the east for the low-alpine zone; above 900 metres in the west and 1200 metres in the east for the middle-alpine zone.

The lapse rate models were then used to identify the mean annual and seasonal 1961 - 1990 isotherm value associated with the upper limit of each zone for both western and eastern mountains, both for mean T_{max} and T_{min}. Following similar reasoning in scenario selection to that described in Section 2.5.3, for this series of experiments the UKCIP02 2050s Medium-Low and 2080s High scenarios were selected to ensure consistency with earlier experiments. Following scenario selection, the ΔT°C changes for the 2050s and 2080s were obtained from HadRM3 Grid 160 and 145 respectively.

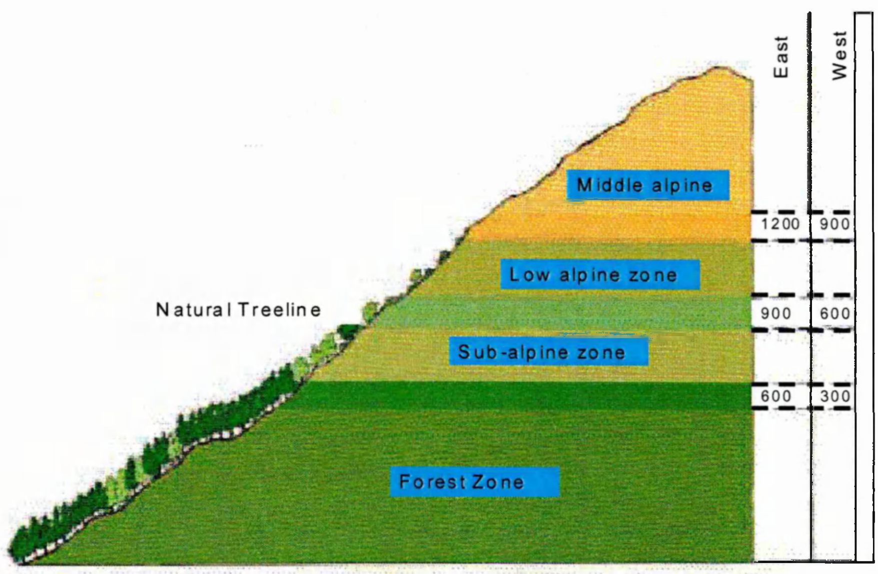


Figure 2.4: Altitudinal vegetation zonations in the Scottish uplands, including approximate altitudes for the different zones in the east and west (adapted from MacKenzie and Gilbert, 2001)

Observed 1961 – 1990 station values for Onich were then deductively adjusted to an elevation of 50 metres using the 10 metre increment lapse rate values. While the ΔT°C output values from HadRM3 Grid 160 for both the 2050s and 2080s were additively adjusted to 50 metres using a combination of the 50 metre and 10 metre lapse rate values. Following the adjustment of station observed and HadRM3 ΔT°C outputs to a common elevation, baseline annual and seasonal values were primed with the adjusted ΔT°C outputs for the 2050s and 2080s and the lapse rate models applied deductively to project the adjusted baseline values to 1300 metres. The same procedures were employed to adjust the Balmoral 1961 -1990 values and HadRM3 Grid 145 $\Delta T^{o}C$ 2050s and 2080s outputs to a common elevation of 300 metres, and the lapse rate models again applied deductively to project the perturbed baseline values to 1300 metres.

Having primed both sets of station baseline values with outputs from the corresponding HadRM3 grid, and having projected shifts altitudinally for both the 2050s and the 2080s. Altitudinal shifts in the isotherm values for both mean minima and mean maxima associated with the upper limit of each of the vegetation zones were noted (nearest 50 metres) for both future time-slices. These shifts were then expressed in metres relative to the 1961 – 1990 isotherm position for the upper limit of each vegetation zone and used to derive the plots for selected seasonal values in Results Sections 5.3.1 and 5.3.2 for Onich and Balmoral respectively.

2.6 Orographically Adjusted Future Precipitation Changes

For purposes of consistency with procedures described in Section 2.4.4.5 and further explored in Sections 4.2.5 and 4.3, areally averaged station data were orographically adjusted to elevations of 500 metres and 1000 metres respectively (Section 5.5). These procedures were applied for both selected seasons (summer and winter) and locations (Section 5.5). For purposes of consistency with the approach elsewhere in this work, these baseline values were perturbed by precipitation changes (Δ%) described by the corresponding HadRM3 grid for the UKCIP02 2050s Medium-Low and 2080s High scenarios (Section 5.4).

3. Assessment Of The 1961-1990 Baseline

Synopsis

The major aims of this Chapter are:

- To consider how baseline climate data may be constructed and quality controlled and used to elucidate some elements of climatic variation across the Highlands.
- Following from this aspects of variability in the station data over the baseline period are then examined. Aspects of variability in the baseline data are then examined in relation to variability in other datasets over a longer time period.
- Finally, linkages are explored between selected seasonal baseline data and fluctuations in NAO Index (NAOI) values.

3.1 Constructing A Baseline - Temperature Stations

3.1.1 Obtaining available baseline years

Following the BADC database search and quality control procedures described in Sections 2.1.2 and 2.1.3, a number of temperature stations with a more or less intact 1961-1990 baseline record were identified. As indicated in Figure 2.2, in terms of geographical spread, available records ranged from the islands of Skye and Rum in the west to Aberdeenshire in the east, with Cape Wrath and Mylnefield representing the most north-westerly and south-easterly of the mainland stations respectively.

However, despite the addition of existing Met Office records to improve the geographical spread of data sources across the region, the number of stations with fully intact records matching IPCC TGCIA guidelines across the region remains sparse (Figure 2.2). This is especially true for much of the Northern Highlands, including Caithness and Sutherland (Figure 2.2). While for Scotland as a whole, where 35% of the land is above 250 m, less than 10% of precipitation gauges are at altitudes above that height (Weston and Roy, 1994). And despite there being in excess of 5000 raingauges in the UK, there are few climatological stations or precipitation gauges sited above 250 m (Taylor, 1976). These difficulties are symptomatic of research problems encountered in mountainous regions generally, where a precise understanding of the climatic characteristics are hampered by a lack of observational data at the spatial and temporal resolution adequate for climate research in regions of complex topography (Beniston, 2000, 2003).

For each of the stations an identification number (Station ID) was allocated and Latitude and Longitude co-ordinates recorded (Table 3.1.1) to facilitate data handling and sorting procedures work described in subsequent sections. The number of intact baseline years was recorded for each station (Table 3.1.2), and specific years not used for each station record over the baseline period noted (Table 3.1.2).

Table 3.1.1: Temperature station records used to construct the 1961-1990 baseline

ID Number	Station Name	Altitude (m)	Latitude (°N)	Longitude (°W)	Baseline Years
1	Aberdeen	52	57.13	-2.14	28
2	Balmoral	283	57.04	-3.22	30
3	Craibstone	102	57.19	-2.21	30
4	Fyvie Castle	55	57.44	-2.39	24
5	Arbroath	29	56.56	-2.57	30
6	Dundee	45	56.47	-2.94	28
7	Montrose	55	56.71	-2.45	30
8	Mylnefield	31	56.46	-3.07	30
9	Fort Augustus	58	57.14	-4.68	26
10	Inverness	4	57.49	-4.22	27
11	Isle of Rhum	5	57.01	-6.28	28
12	Onich	15	56.72	-5.22	28
13	Prabost	67	57.47	-6.31	28
14	Dall, Rannoch	232	56.68	-4.30	23
15	Faskally	94	56.72	-3.77	30
16	Fortrose	5	57.57	-4.09	25
17	Kinlochewe	25	57.61	-5.31	24
18	Poolewe	6	57.77	-5.60	27
19	Cape Wrath	112	58.63	-5.00	30
20	Tiree	12	56.50	-6.88	Met Office Figures
21	Ardtalnaig	130	56.53	-4.11	Met Office Figures
22	Braemar	339	57.01	-3.40	Met Office Figures
23	Kinloss	5	57.65	-3.56	Met Office Figures
24	Stornoway	15	58.21	-6.32	Met Office Figures
25	Kirkwall	26	58.95	-2.90	Met Office Figures

From Tables 3.1.1 and 3.1.2 it is apparent that due to missing data for some years and substantial gaps due to logger/instrument failure in others, intact baseline years available range from 23-30 years (27.68 mean intact record years for the nineteen stations). Table 3.1.2 also indicates that there is no consistent pattern in station record gaps, although the 1960s and 1970s seem to have been more problematic decades for some of the station records (Figure 3.1.1).

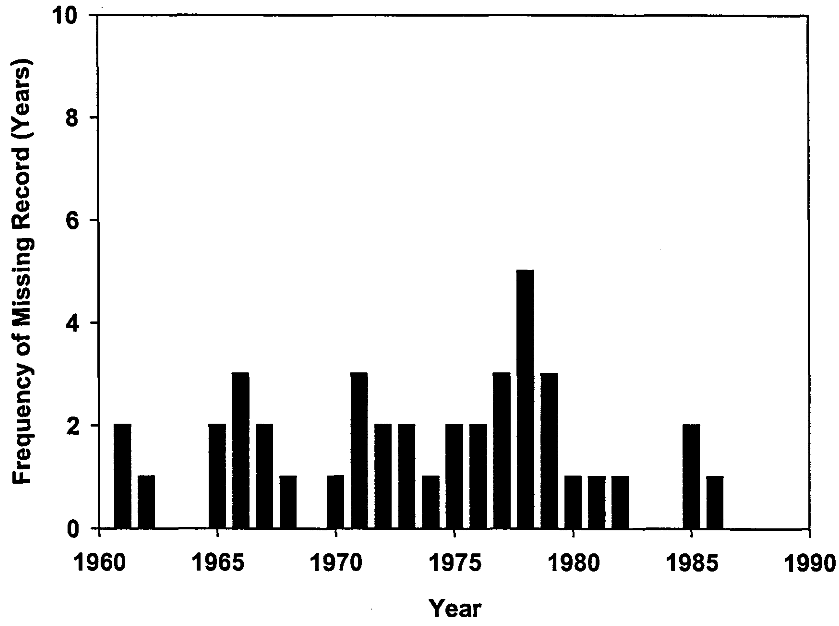


Figure 3.1.1.: Frequency distribution of missing year records (all stations) 1961-1990

Despite these problems, annual and seasonal values of T_{max} and T_{min} for the 19 station records obtained via the BADC were calculated for the available years as per the procedures described in Sections 2.1.1 – 2.1.3. These are presented together with the values obtained from the existing Met Office records in Appendix Tables 3.1 (T_{max}) and 3.2 (T_{min}).

Table 3.1.2: Years missing data for the nineteen BADC station records

ID Number	Station Name	Years Not Used
1	Aberdeen	1977, 1978
2	Balmoral	0
3	Craibstone	0
4	Fyvie Castle	1965, 1966, 1971, 1973, 1985, 1986
5	Arbroath	0
6	Dundee	1976, 1989
7	Montrose	0
8	Mylnefield	0
9	Fort Augustus	1971, 1972, 1974, 1979
10	Inverness	1975, 1977, 1978
11	Isle of Rhum	1977, 1978
12	Onich	1966, 1972
13	Prabost	1989, 1990
14	Rannoch School	1965, 1966, 1967, 1968, 1970, 1973, 1976 1976
15	Faskally	0
16	Fortrose	1961, 1971, 1978, 1979, 1985
17	Kinlochewe	1975, 1978, 1979, 1980, 1981, 1982
18	Poolewe	1961, 1962, 1967
19	Cape Wrath	0

The mixture of maritime and continental influences on Highland climate are evident in many of the mean annual and seasonal station values for the 1961-1990 baseline period (Appendix Tables 3.1 & 3.2). To illustrate this, from the data set presented in Apppendix Tables 3.1 and 3.2, the annual and seasonal range of T_{mean} ($[T_{max} + T_{min}]$ / 2) was calculated for a representative subset of the data (Figure 3.1.2). This transect based approach for representative four stations west to east is used to illustrate and summarise some of these features of the regional climate (Figure 3.1.2).

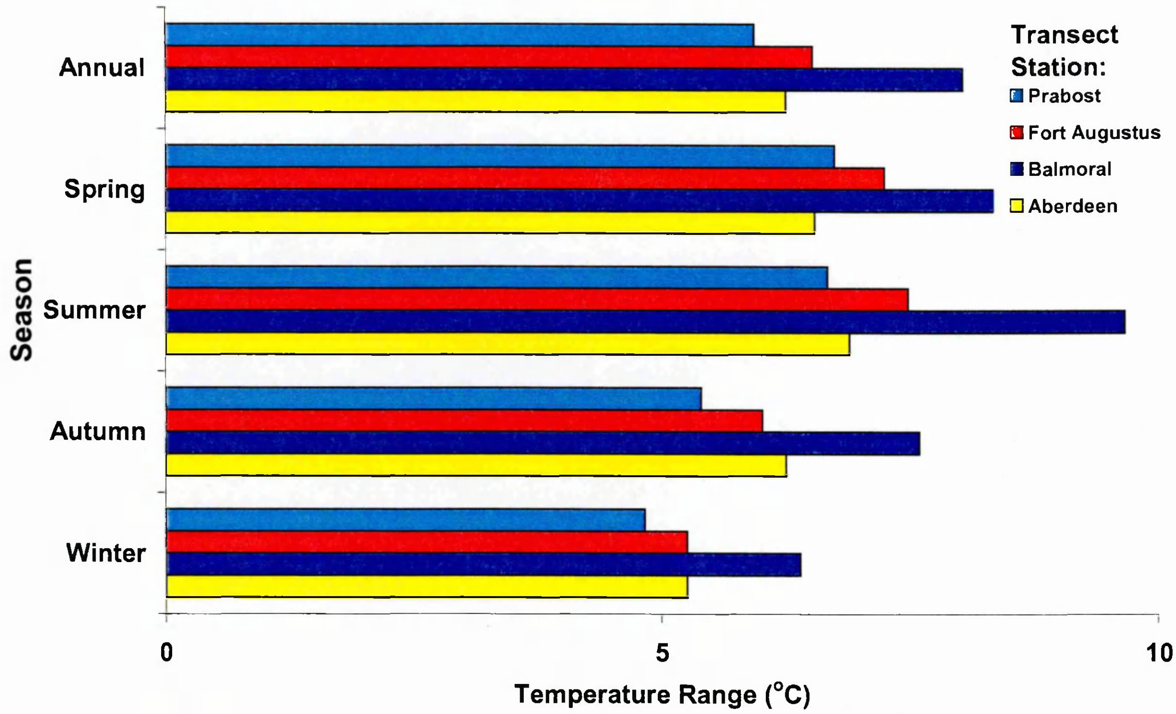


Figure 3.1.2: Mean annual and seasonal temperature range ($T_{max} - T_{min}$), four stations west to east for the 1961-1990 baseline period. Note: T_{mean} annual and seasonal values ($[T_{max} + T_{min}]/2$) are the infilled baseline values obtained following the statistical procedures described in Section 3.3 below.

Therefore, for stations in the more maritime influenced west such as Tiree, Prabost, Poolewe and the Isle of Rum, there is a marked suppression in the mean annual and seasonal temperature range recorded for the baseline period. This contrasts with stations further inland and to the east where the mix of oceanic and continental influences is much more evident in the baseline data. Figure 3.1.2 helps illustrate this, thus the markedly more continental influences on the climate of Balmoral by comparison with Prabost on Skye is evident. Whereas these continental influences are not quite so marked for a site such as Fort Augustus in a more central inland location, although the annual and seasonal range of temperature here is still greater than for west coast sites such as Prabost.

When compared with Balmoral in an eastern inland location, the continental influences at Aberdeen are not quite so marked due to coastal effects and the contrasting topographic setting. However, the maritime influence is not as marked at eastern coastal sites by comparison with more oceanic sites to the west such as Prabost. Superimposed on the contrasting atmospheric circulation influences, these site differences also reflect the contrasting thermal properties and mixing characteristics of the North Sea as a relatively enclosed basin system compared to the more oceanic Atlantic waters supplied by the North Atlantic Drift.

While many of these broad features of regional climate are readily identifiable from the baseline mean values, at a localised scale the effects of surrounding topography can also be identified in a number of the station records. For instance, Kinlochewe in the oceanic west is a notably 'cold' station in terms of winter T_{min} (0.54°C), by comparison with the Isle of Rum (2.20°C) and Poolewe (2.10°C) (Appendix Table 3.2). While not as markedly low as at Kinlochewe, winter T_{min} values for both Onich (1.03°C) and Prabost (1.35°C) also appear to be affected by the surrounding topography. In common with Kinlochewe these are sites adjacent to relatively high hills, and it would seem winter T_{min} values are reflecting localised frost hollow effects under certain atmospheric conditions over the long baseline period.

Similar effects are also evident in the winter T_{min} values associated with stations further inland. For example, Rannoch School (-1.69C) and Faskally (-0.66°C) are notably cold, while frost hollow effects also seem to affect winter T_{min} values at Fort Augustus (0.26°C) and Ardtalnaig (0.60°C), although to a lesser extent. However, none of these stations are as notably cold as Balmoral (-2.01°C) and Braemar (-

1.97°C) further to the east. Here a combination of the surrounding high hills and local valley topography lead to a ponding of cold air under certain conditions, and for these locations the suppression of spring T_{min} values is also evident, (0.96°C) and (1.10°C) at Balmoral and Braemar respectively . These particular site characteristics are well known, for example Jones and Lister (2004a, b) found Braemar to be the coolest station in the long records they analysed.

This combination of station elevation (283m and 339m respectively for Balmoral and Braemar) and surrounding high hills appears to be a significant control on spring temperatures, as Rannoch School (232m) in an analogous topographic location also records a low spring T_{min} (1.03°C). In their analysis, Jones and Lister (2004,a, b) found outlying cold anomalies detected at Braemar were not present at the coastal sites used in their study. These were related to extensive snow-lie in the winter half-year and hence were accepted in their homogenised record (Jones and Lister 2004,a, b).

Such anomalies are to be expected at this site as Braemar is also noted for its long record documenting periods of snow-lie. In a site-specific interpretation of the record, it is therefore questionable if 'anomaly' is an appropriate moniker for outlying cold values. However, since the record also exhibits considerable inter-annual variation in the number of days with snow lying (Harrison *et al.*, 2001), apparently anomalous years in the record linked to the presence of late-lying snow should be expected.

While all of the above are well documented features of local climate in the Highlands, they are worth re-iterating in the context of this work. Subsequent sections conducting an initial evaluation of HadRM3's performance for the region will return to some of these localised controls on climate and explore them further in the context of climate model performance for the region.

3.2 Constructing A Baseline - Precipitation Stations

3.2.1 Obtaining available baseline years

Procedures here were largely analogous to those described for temperature in Section 3.1. However, in this instance the data searching and handling protocols described in Section 2.2 were followed. In terms of spatial cover (Figure 2.2), the available records ranged from Rum and Skye in the west to the Angus glens in the east. Audit procedures on data availability again revealed that much of the region remains data sparse in terms of a reasonably intact 1961-1990 baseline record, particularly in the north.

As for temperature and for the same reasons, a station ID was allocated and Latitude and Longitude co-ordinates and intact baseline years recorded for each station (Tables 3.2.1 and 3.2.2). Specific years missing for each station record were also noted (Tables 3.2.3 and 3.2.4). Apparent from both Tables 3.2.1 and 3.2.2 is that intact baseline years range from 21-29 years (27.58 mean intact record years for the fifty five stations), and that the 1960s seem a particularly problematic decade in terms of missing data (Figure 3.2.1) . Despite these problems, annual and seasonal values of mean total precipitation for the 55 station records were again calculated for the available years as per the procedures described in Sections 2.2.2 and 2.2.3 and are recorded in Appendix Table 3.5.

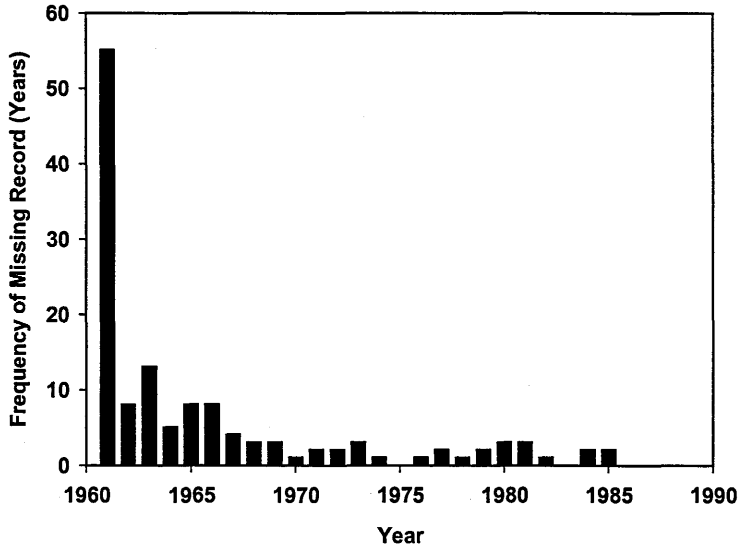


Figure 3.2.1: Frequency distribution of missing year records (all stations) 1961-1990

It is worth noting in the context of the interpretation previously offered in Section 2.2.3 (drifting snow affecting funnel gauges), that the biggest gap year across the station records is coincident with the unusually cold winter of 1961/1962, and that these gaps continue into the exceptionally snowy and cold winter of 1962/1963, although not to the same extent (Figure 3.2.1; Tables 3.2.1 & 3.2.2). March 1962 was the coldest March of the 20th century in the Central England Temperature (CET) record and the coldest since 1892 (Lamb, 1982) with a mean monthly temperature of 2.8°C. While, coincident with continental easterly winds, November 1962 was unusually snowy with level snowlie of 17 cm in parts of Scotland, and with drifts of up to ~1 metre (Lamb, 1982). The last week of December in the same year was very snowy and preceded the 'great winter' of 1962/1963, the coldest of the 20th century in the CET and the second coldest since 1739-1740 in the entire series (Lamb, 1982; Wright 1976). Associated with this, the CET recorded January and February mean temperatures of –2.1°C and –0.7°C respectively for 1963.

Table 3.2.1: Precipitation records used to construct the 1961-1990 baseline

ID Number	Station Name	Altitude (m)	Latitude (°N)	Longitude (°W)	Baseline Years
1	Lumphanan	145	57.12	2.70	27
2	Balnakeilly	209	56.68	3.17	29
3	Barney	345	56.76	3.22	29
4	Clintlaw	317	56.74	3.21	29
5	Glen Lee	471	56.92	3.03	29
6	Glenhead	366	56.77	3.19	29
7	Longdrum	319	56.74	3.18	29
8	Newton	224	56.70	3.17	29
9	The Linns	396	56.82	3.31	29
10	Westerton	274	56.72	3.17	29
11	Allt Leachdach	198	56.87	4.86	29
12	Alltbeithe	242	57.11	5.27	28
13	Fort William	66	56.81	5.11	25
14	Glen Einich	475	57.10	3.77	22
15	Glenfeshie Lodge	357	57.02	3.90	25
16	Loch a' Chrathaich	536	57.25	4.71	28
17	Loch Eilde Mhor	343	56.73	4.91	29
18	Loch Pattack	433	56.89	4.39	29
19	Lochan na Sgud	143	57.04	5.20	29
20	Lon Mor	160	57.12	4.75	24
21	Rhum, Coire Dubh	320	57.00	6.30	29
22	Rhum, Harris	21	56.98	6.38	28
23	Rhum, Kilmory	15	57.05	6.36	29
24	Rhum, Slugan Burn	152	57.00	6.29	29
25	Skye, Alltdearg House	55	57.28	6.19	29
26	Sronlairig Lodge	358	57.13	4.42	27
27	Tromie Dam	436	56.97	4.03	24
28	Allt Dubhaig	416	56.83	4.23	28

Table 3.2.1 (contd.): Precipitation records used to construct the 1961-1990 baseline

ID Number	Station Name	Altitude (m)	Latitude (°N)	Longitude (°W)	Baseline Years
29	Allt Girnaig	259	56.74	3.76	27
30	Allt na Bogair	290	56.65	4.29	26
31	Ben Vorlich	433	56.32	4.19	28
32	Bruar Intake	381	56.83	3.93	27
33	Capel Hill North	411	56.66	3.57	29
34	Capel Hill West	354	56.65	3.58	29
35	Chapel Burn	308	56.27	3.52	29
36	Frandy Burn	319	56.21	3.72	29
37	Loch Benally South	323	56.62	3.50	29
38	Loch Ordie	299	56.63	3.58	29
39	Lochan Oissineach Mhor	427	56.67	3.59	29
40	Slateford Burn	236	56.27	3.47	28
41	Sronphadruig Lodge	497	56.88	4.11	29
42	Am Meallan	229	57.35	5.53	29
43	Carn Anthony	107	57.57	5.38	24
44	Coulin Pass	290	57.50	5.30	25
45	Gleann Fhiodhaig	223	57.49	5.04	27
46	Glenshiel Forest	183	57.17	5.28	23
47	Grudie Bridge	27	57.65	5.39	21
48	Heights of Kinlochewe	25	57.62	5.23	27
49	Loch Fannich South	293	57.63	5.05	28
50	Loch Glass	219	57.70	4.46	29
51	Loch Gleann na Muice	316	57.66	5.25	27
52	Luibachlaggan	259	57.76	4.77	28
53	Strathrannoch	290	57.73	4.71	29
54	Strone Nea	366	57.81	5.02	27
55	Tornapress	98	57.25	5.59	26

Table 3.2.2: Years missing data for the BADC station records

ID Number	Station Name	Years Not Used
1	Lumphanan	1961, 1964, 1965
2	Balnakeilly	1961
3	Barney	1961
4	Clintlaw	1961
5	Glen Lee	1961
6	Glenhead	1961
7	Longdrum	1961
8	Newton	1961
9	The Linns	1961
10	Westerton	1961
11	Allt Leachdach	1961
12	Alltbeithe	1961, 1963
13	Fort William	1961, 1962, 1963, 1964, 1965
14	Glen Einich	1961, 1962, 1963, 1966, 1967, 1973, 1984, 1985
15	Glenfeshie Lodge	1961, 1965, 1966, 1976, 1977
16	Loch a' Chrathaich	1961, 1985
17	Loch Eilde Mhor	1961
18	Loch Pattack	1961
19	Lochan na Sgud	1961
20	Lon Mor	1961, 1977, 1978, 1979, 1980, 1981
21	Rhum, Coire Dubh	1961
22	Rhum, Harris	1961, 1963
23	Rhum, Kilmory	1961
24	Rhum, Slugan Burn	1961
25	Skye, Alltdearg House	1961
26	Sronlairig Lodge	1961, 1962, 1963
27	Tromie Dam	1961, 1962, 1963, 1964, 1965, 1966
28	Allt Dubhaig	1961, 1965

Table 3.2.2 (contd.): Years missing data for the BADC station records

ID Number	Station Name	Years Not Used
29	Allt Girnaig	1961, 1962, 1963
30	Allt na Bogair	1961, 1965, 1966, 1968
31	Ben Vorlich	1961, 1969
32	Bruar Intake	1961, 1963, 1964
33	Capel Hill North	1961
34	Capel Hill West	1961
35	Chapel Burn	1961
36	Frandy Burn	1961
37	Loch Benally South	1961
38	Loch Ordie	1961
39	Lochan Oissineach Mhor	1961
40	Slateford Burn	1961, 1965
41	Sronphadruig Lodge	1961
42	Am Meallan	1961
43	Carn Anthony	1961, 1966, 1973, 1980, 1981, 1982
44	Coulin Pass	1961, 1963, 1979, 1980, 1981
45	Gleann Fhiodhaig	1961, 1971, 1972
46	Glenshiel Forest	1961, 1962, 1963, 1964, 1965, 1966, 1969
47	Grudie Bridge	1961, 1966, 1967, 1969, 1970, 1971, 1972, 1973, 1974
48	Heights of Kinlochewe	1961, 1967, 1968
49	Loch Fannich South	1961, 1984
50	Loch Glass	1961
51	Loch Gleann na Muice	1961, 1967, 1968
52	Luibachlaggan	1961, 1963
53	Strathrannoch	1961
54	Strone Nea	1961, 1962, 1963
55	Tornapress	1961, 1962, 1963, 1966

As was noted for temperature, the maritime influence on Highland climate and the profound influence of the predominantly westerly weather systems is apparent for many of the station records in Appendix Table 3.5., particularly for locations in the west which experience much higher Precip totals by comparison with more easterly sites in the lee of the western mountains. A transect based approach is used again to demonstrate the spatial pattern of climate across Highland Scotland, and, in particular, to illustrate the profound influence of topography and orography on Precip distribution totals (Figure 3.2.2).

In this case, rather than using individual station records to illustrate the west-east rainfall gradient, the regionally averaged Precip station data described in Section 3.4.2 below were used to construct the mean annual and seasonal plots. It is worth noting that no statistical infilling methods have been applied to the regionally averaged Precip station records used to construct Figure 3.2.2.

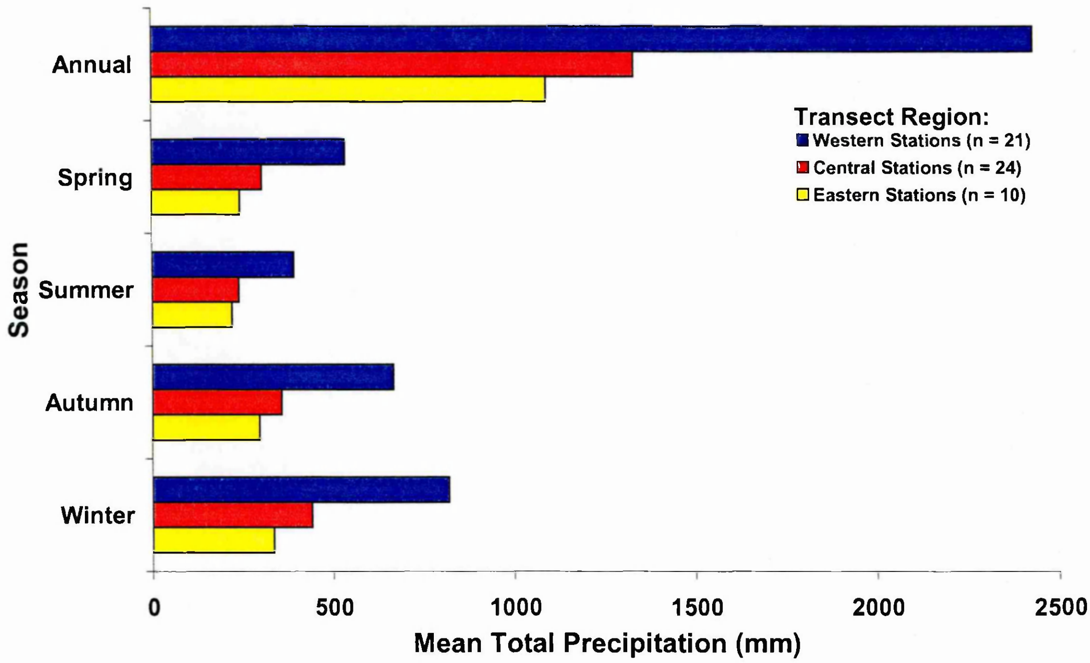


Figure 3.2.2: Mean total precipitation (Precip) for regionally averaged 1961-1990 baseline station data

Further information on the station groupings (including the latitudinal and longitudinal ranges) can be found in Section 3.4.2. Harrison (1997), and Harrison and Kirkpatrick (2001) report on the dominant role of topography in shaping the spatial pattern of rainfall for the region. Figure 3.2.1 supports this observation and illustrates the well known west to east rainfall gradient across the region. The station data also illustrate another notable feature of Highland climate, the effectiveness of the north-south barrier which the mountains of the Western Highland present to Atlantic winds from the south-west and north-west (Roy, 1997).

Significant reductions in total Precip both annually and seasonally are apparent for stations experiencing rain shadow effects in the lee of the western mountains, with a clear west-east gradient obvious. Since the monthly Precip station data used spanned a range of altitudes (Section 3.4.2) and encompassed a range of aspects. It would seem the regional averaging exercise smoothed out most of the local variation attributable to station siting and usefully captured this broad feature of Highland climate for the 1961-1990 baseline period.

3.3 Statistical Procedures for Infilling Missing Station Data

3.3.1 Temperature

As discussed in Section 2.4.4.2, some method of infilling missing data in the 1961-1990 baseline was required in order to improve confidence in the findings of some aspects of subsequent work (Chapters 4 and 5). With cross-correlations between all nineteen BADC station records identified as the preferred method, Pearson Correlation Coefficients (r-values) and their associated p-values obtained from the larger output matrix (for station data sets requiring this treatment) are summarised in Table 3.3.1.

Table 3.3.1: Pearson correlation coefficient (r) and p-values from correlation matrices (n = 19) for best fit stations, annual T_{max} and T_{min}

Station Record Requiring Infilling	**Corresponding Station Record For T_{max}**	**Pearson Correlation Coefficient (r) and p Values**	**Corresponding Station Record For T_{min}**	**Pearson Correlation Coefficient (r) and p- values**
Aberdeen	Craibstone	0.983, $p < 0.001$	Craibstone	0.892, $p < 0.001$
Fyvie Castle	Craibstone	0.955, $p < 0.001$	Balmoral	0.853, $p < 0.001$
Dundee	Arbroath	0.951, $p < 0.001$	Arbroath	0.928, $p < 0.001$
Fort Augustus	Balmoral	0.897, $p < 0.001$	Inverness	0.824, $p < 0.001$
Inverness	Balmoral	0.963, $p < 0.001$	Dundee	0.871, $p < 0.001$
Isle of Rhum	Mylnefield	0.841, $p < 0.001$	Dundee	0.786, $p < 0.001$
Onich	Isle of Rhum	0.798, $p < 0.001$	Isle of Rhum	0.888, $p < 0.001$
Prabost	Isle of Rhum	0.960, $p < 0.001$	Isle of Rhum	0.930, $p < 0.001$
Rannoch School	Inverness	0.959, $p < 0.001$	Onich [1]	0.892, $p < 0.001$
Fortrose	Mylnefield	0.725, $p < 0.001$	Craibstone	0.533, $p = 0.006$
Kinlochewe	Isle of Rhum [2]	0.942, $p < 0.001$	Fort Augustus	0.851, $p < 0.001$
Poolewe	Prabost	0.971, $p < 0.001$	Prabost	0.865, $p < 0.001$

Notes:
1. Onich missing 1966; Prabost ($r = 0.809$; $p < 0.001$) record used.
2. Isle of Rum missing 1978; Faskally ($r = 0.933$; $p < 0.001$) record used.

For identified stations with missing records, the corresponding stations (for record infilling purposes) with the highest correlation coefficient and associated r and p-values were recorded following the procedures described in Section 2.4.4.2. Given the variably missing years across individual station records for the baseline period, not all of the missing data could be infilled from the most closely correlated station records. As a result, in a few cases the next 'best fit' station data record from matrix outputs were utilised. These are numerically annotated in Table 3.3.1.

Table 3.3.1 summarises correlation data for stations where infilling of values was required. Correlations are based on all of the nineteen BADC station records and for all existing annual values of T_{max} and T_{min} for the 1961-1990 record. Since the existing Met Office station data used to extend spatial cover only provided 30-year mean values, these could not be used in this procedure. Therefore for the nineteen baseline station records 525 values of a possible 570 were used in total, i.e. 19 stations x 30 baseline years (570 annual values) minus the sum of all missing years across the baseline station record (45).

Correlations obtained between all nineteen station data records used in the matrices were generally high (T_{max}, $r = 0.725$-0.960, $p < 0.001$; T_{min}, $r = 0.533$-0.930, $p < 0.001$ - 0.006). This most likely reflects the regional context and the maritime influence in subduing the overall amplitude of temperature variation by comparison with more continental climates elsewhere. However, even in the local scale context of this work, correlations were generally highest between adjacent geographic locations (e.g. Aberdeen and Craibstone, Poolewe and Prabost), or for stations in analagous topographic locations (e.g. Fort Augustus and Balmoral, Kinlochewe and Fort Augustus). See also preceding commentary in Section 3.1.1 on site influences on the local thermal climate. Although it is unusual to find that the Mylnefield station annual T_{max} values on the east coast should be most closely correlated to those of the Isle of Rhum in the west and hence used for infilling the missing baseline data.

The rigour of this exercise could have been improved had the above procedures been repeated for each season, or better, by month for each station across the thirty year duration of the baseline period and these more refined values used to infill missing data records. However, such an exercise would have been both time consuming and computationally complex (particularly for Precip records, Section 3.2.2) and beyond the scope of this study. Consequently, while not ideal, the approach adopted was deemed as the 'best practicable option'.

Following these statistical procedures, station record data from the most closely correlated stations (Table 3.3.1) were used to infill the missing years previously identified in Table 3.1.2 for the 30-year baseline period. These adjusted (infilled) 1961-1990 values are recorded in Appendix Tables 3.3 (T_{max}) and 3.4 (T_{min}) for stations where the procedure was applied.

Examination of the differences obtained for the adjusted versus unadjusted baseline data support some of the above qualifications in terms of the approach. If differences in the T_{max} values provided in Appendix Tables 3.3 and 3.4 are quantified (adusted [infilled] baseline – unadjusted [non-infilled] baseline), the absolute changes to mean annual and seasonal values associated with the infilling exercise to replace missing years in the record are not that substantial. For example, changes to annual T_{max} values are in the range –0.14°C (Inverness and Fort Augustus) - +0.21°C (Rannoch School). While the adjusted T_{min} values span a wider range than those for T_{max}, again the absolute differences in values are not that substantial (Appendice Tables 3.2 and 3.4). Using the annual mean example again (T_{min}), differences range from –0.3°C (Fyvie Castle) - +0.55 °C (Rannoch School).

3.3.2 Precipitation

Most of the considerations for precipitation were largely similar to those described for temperature in 3.1.1 and correlation procedures were identical. Given the problems with variably missing years across the record previously noted for temperature, numerically annotated notes beneath Table 3.3.2 again indicate instances where station records other than those tabulated were used to infill missing data.

Table 3.3.2: Pearson correlation coefficient (r) and p-values from correlation matrices (n = 55) for best fit stations, annual total Precip

Station Record Requiring Infilling	Corresponding Station Record For Annual Precip	Pearson Correlation Coefficient (r) and p-values
Lumphanan	Balnakeilly	0.830, p <0.001
Alltbeithe	Allt Leachdach	0.893, p <0.001
Fort William	Allt Leachdach	0.901, p <0.001
Glen Einich	Glenfeshie Lodge	0.846, p <0.001
Glenfeshie Lodge	Glen Einich [1]	0.846, p <0.001
Loch a' Chrathaich	Fort William	0.894, p <0.001
Lon Mor	Fort William	0.952, p <0.001
Rhum, Harris	Rhum, Coire Dubh	0.897, p <0.001
Sronlairig Lodge	Loch a' Chrathaich	0.890, p <0.001
Tromie Dam	Glenfeshie Lodge [1,2]	0.927, p <0.001
Allt Dubhaig	Loch Pattack	0.893, p <0.001
Allt Girnaig	Newton	0.837, p <0.001
Allt na Bogair	Allt Dubhaig [3]	0.939, p <0.001
Ben Vorlich	Glenfeshie Lodge	0.829, p <0.001
Bruar Intake	Glenfeshie Lodge	0.883, p <0.001
Slateford Burn	Chapel Burn	0.954, p <0.001
Carn Anthony	Lon Mor [4]	0.895, p <0.001
Coulin Pass	Lon Mor	0.921, p <0.001
Gleann Fhiodhaig	Lon Mor	0.933, p <0.001
Glenshiel Forest	Rhum, Coire Dubh	0.720, p <0.001
Grudie Bridge	Alltbeithe	0.941, p <0.001
Heights of Kinlochewe	Grudie Bridge [5]	0.953, p <0.001
Loch Fannich South	Lon Mor	0.960, p <0.001
Loch Gleann na Muice	Carn Anthony	0.920, p <0.001
Luibachlaggan	Carn Anthony	0.816, p <0.001
Strone Nea	Grudie Bridge	0.939, p <0.001
Tornapress	Lon Mor	0.934, p <0.001

Notes:
1. Glenfeshie Lodge & Glen Einich both missing 1966; Fort William (r = 0.742, p <0.001) record used.: **2.** Tromie Dam & Glenfeshie Lodge both missing 1965; Sronlairig Lodge (r = 0.864, p <0.001) record used. : **3.** Allt na Bogair & Allt Dubhaig both missing 1965; Loch Pattack (r = 0.916, p <0.001) record used.: **4.** Carn Anthony & Lon Mor both missing 1980; Am Meallan (r = 0.877, p <0.001) record used.: **5.** Heights of Kinlochewe & Grudie Bridge both missing 1967; Lon Mor (r = 0.910, p <0.001) record used

Table 3.3.2 summarises correlation coefficients values for stations which required infilled values. Correlations were based on the fifty five station records and for all existing annual values of Precip for the 1961-1990 period. Therefore 1, 517 values of a possible 1, 650 were used in total; 55 stations x 30 baseline years (1, 650 annual values) minus the sum of all missing years for the baseline (133).

Correlation coefficients and their associated p-values for the fifty five station datasets had considerably more range than for those obtained between the temperature datasets (Precip, $r = -0.675$ - $+0.981$; $p <0.001$ - $= 0.999$). Given the much wider range of precipitation station locations, elevations and aspects, this greater variability in association (dis-association) between datasets is not surprising given the highly localised controls of orography, aspect and slope which govern variability in precipitation. Most of the low to negative correlations with statistically insignificant p-values were obtained for stations in contrasting western and eastern settings. Given the contrasting precipitation regimes east to west for the baseline period highlighted in Figure 3.2.1 such poor or spurious correlations should be expected.

However, and as also would be expected, the higher correlation coefficients and their associated p-values ($r = 0.720 - 0.960$; $p = <0.001$) presented in Table 3.3.2 above are obtained between stations in largely analogous geographic and topographic settings. To illustrate this, the scatter plot in Figure 3.3.2.1 shows the relatively close correspondence in longitude for the stations referred to in Table 3.3.2 when they are plotted against each other. Given the dominant control of the west-east precipitation gradient across the region it should be expected that correlations between station datasets will reflect this. Regarding the overall rigour of the approach (using annual

Precip totals only) and scope for improvement, the same general observations made for temperature also apply here (see Section 3.3.1).

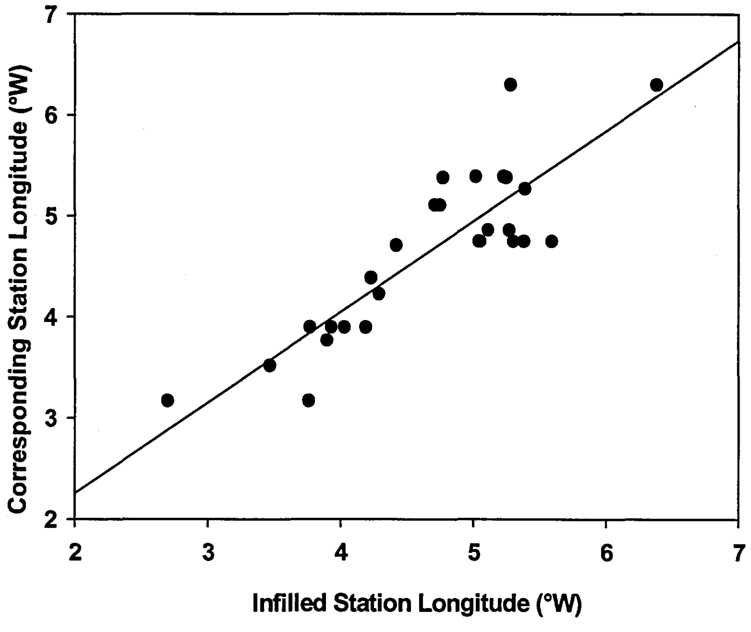


Figure 3.3.2.1: Infilled station longitude co-ordinates plotted against corresponding station longitude co-ordinates for stations referred to in Table 3.3.2

As with temperature, examination of the differences obtained for the adjusted versus unadjusted baseline precipitation data support some of the qualifications to the approach of using only annual mean totals for the correlations. When Precip differences between Appendix Tables 3.5 & 3.6 are quantified (adusted [infilled] baseline – unadjusted [non-infilled] baseline), the absolute changes to mean annual and seasonal totals are not that great. For example, changes to annual values associated with the data infilling exercise encompass a range of –211.06mm (Carn Anthony) - + 219.84mm (Grudie Bridge).

3.4 Variability In Station Data Over the 1961-1990 Baseline and Relationships to Variability in Longer Time-Series

3.4.1 Variability in T_{max} and T_{min}

Following the procedures described in Section 2.3 anomaly plots with respect to the 1961-1990 regional mean were generated for selected values. While this exercise was conducted for the full seasonal range of maxima and minima, only mean annual T_{max} and T_{min} plots are presented here since these can be most readily related to trends in datasets with more extensive spatial coverage and on longer time-series available from elsewhere. The Hadley Centre/CRU dataset details have been described previously in Section 2.3. Later in this section variability in the 1961-1990 baseline data will also be examined in relation to the Scottish and Northern Ireland Forum for Environmental Research (SNIFFER) dataset. Earlier work on these long time-series by Purves *et al.* (2000) has been more recently re-visited by Jones and Lister (2004, a, b).

Unlike the procedures described in 3.3 and as was noted in Section 2.3, no statistical procedures were used to infill missing station values. Available intact year records 1961-1990 for each station were simply summed and averaged, with corrections being made to divisions to account for missing baseline years for each of the stations used to construct the regional average. Part of the reasoning here was that for stations in a largely similar climatic zone, regional averaging of the datasets would tend to smooth out the distortion of individual stations missing data for certain years. Similarly, any differences due to station siting characteristics such as elevation and aspect would be incorporated into any regional pattern of variability.

Provisional classification of stations into groups was undertaken to be coincident with some more objectively mapped climatic zonations such as the Conrad Index (CI) of continentality previously discussed and defined in Section 1.3.4 (Figure 1.7). While station groupings were largely provisional and dependant on the occurrence of baseline data availability, underpinning the selection was a desire to capture some of the gradation of climate across the region from the more oceanically influenced west to the more continentally influenced east, and to elucidate any possible differences in the amplitude of variation and trends in the data.

However, some aspects of the exercise perforce remained subjective. For instance Kinlochewe data was not included for the Region 1 stations since it was missing more years (six) by comparison with the five stations used (Table 3.1.2). Whereas Rannoch School data with seven missing years was included for the Region 2 station groupings in order to improve spatial cover in the central inland section of the region. This rather arbitrary grouping of stations is apparent from Figure 3.4.1. Precipitation station groupings are also indicated here for convenience and to avoid figure duplication, but see Section 3.4.2 for discussion.

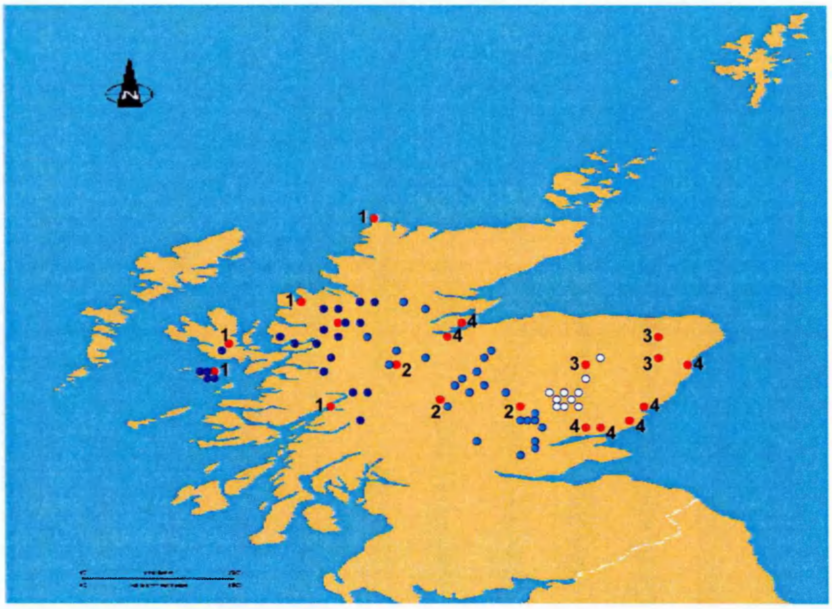


Figure 3.4.1: Outline map indicating station groupings for regional averaging. Temperature stations in red, numbers denote regional groupings (see text). Region 1 precipitation stations are in dark blue, Region 2 in light blue and region 3 in white.

Station groupings indicated in Figure 3.4.1 are loosely linked with some of the CI zones as below;

- Isle of Rhum, Prabost, Onich, Poolewe and Cape Wrath (Region 1 stations; anomaly plots Figure 3.4.2) are broadly coincident with the hyper-oceanic/oceanic west and relate to CI scores of 2-6. For this station group (n = 5), latitude and longitude ranged from 56.72° - 58.63° and -6.31° – -5.0° respectively, with an elevation range of 5m – 67m.
- Fort Augustus, Rannoch School and Faskally (Region 2 stations; anomaly plots Figure 3.4.2) are coincident with the drier inland zone recording CI scores of 6-10, but with a more central bias. For this station group (n = 3), latitude and longitude ranged from 56.68° – 57.14° and -4.68° - -3.77° respectively, with an elevation range of 58m – 232m.

- Balmoral, Fyvie Castle and Craibstone (Region 3 stations; anomaly plots Figure 3.4.2) are coincident with the same CI zone, but with a more eastern bias. For this station group (n = 3), latitudinal and longitudinal ranges are 57.04° – 57.44° and – 3.22° - -2.21° respectively, with an elevation range of 55m – 283m.
- Fortrose, Inverness, Mylnefield, Aberdeen, Arbroath, Dundee and Montrose (Region 4 stations; Figure 3.4.2) correspond to the continentally influenced eastern coastal zones with CI score of 6-8. For this station group (n =7), latitude and longitude ranged from 56.46 – 57.57 and –4.22 - -2.14 respectively, with an elevation range of 4m – 55m.

The T_{max} and T_{min} anomalies for the Region 1 stations reveal a fluctuating pattern of warming and cooling throughout the 1960s and 1970s (Figure 3.4.2). The cold year of 1979 is notable, with further temperature fluctuations throughout the 1980s, including an obvious mid-decade cooling.

T_{max} and T_{min} anomalies for the Region 2 stations show a broadly similar pattern and again he cold year of 1979 is again notable, as is the 1980s mid-decade cooling (Figure 3.4.2). However, as would be expected for less maritime-influenced inland locations, the amplitude of variation (particularly for T_{min}) is considerably more marked compared to the western stations.

Patterns of variation in T_{max} and T_{min} anomalies for the Region 3 and 4 stations are also broadly similar (Figure 3.4.2) for episodes of warming and cooling over the decades of the baseline period. However, for these more-continentally influenced eastern locations, the overall amplitude of variation (particularly for minima) varies compared to the more maritime-influenced western stations.

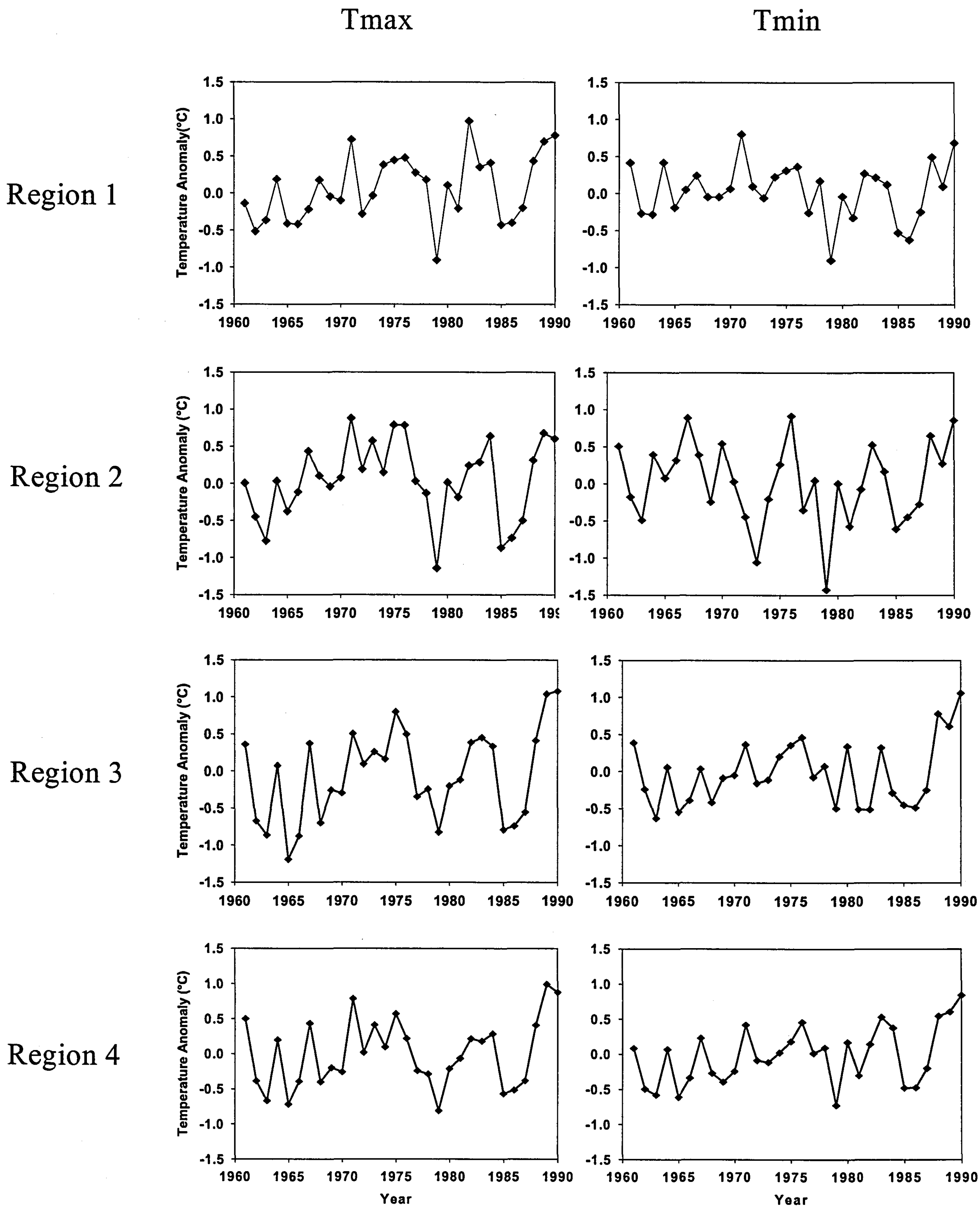


Figure 3.4.2: Regionally averaged annual anomalies 1961-1990 for Region 1, 2,3 and 4 stations T_{max} and b, T_{min}

The notable late 1970s cooling evident in all the records is consistent with trends elsewhere in Europe, where relatively cool temperatures prevailed in the 1970s

(Butler *et al.*, 2005). Trend analysis of fluctuations in mean annual T_{max} and T_{min} (as opposed to anomalies) for the baseline period reveals the consistency of this 1970s cooling across the Highlands, as well as providing some quantification of trends in other decades for the baseline period (Table 3.4.1).

Table 3.4.1: Annual temperature trends for mean T_{max} and T_{min} by region and decade for the 1961-1990 baseline period

Temperature Trends By Decade and Across The Baseline (°C)				
Station Region	**1960s**	**1970s**	**1980s**	**1961-1990**
1 T_{max}	+0.03	-0.06	+0.04	+0.02
T_{min}	-0.004	-0.09	+0.05	-0.003
2 T_{max}	+0.06	-0.13	+0.04	+0.006
T_{min}	+0.04	-0.008	+0.09	-0.003
3 T_{max}	-0.006	-0.10	+0.08	+0.02
T_{min}	-0.01	-0.01	+0.15	+0.01
4 T_{max}	-0.02	-0.12	+0.08	+0.01
T_{min}	+0.001	-0.03	+0.08	+0.02

Table 3.4.1 also reveals the contrasting pattern for the 1960s when western and central stations (Region 1 and Region 2) are compared with those further to the east (Region 3 and 4). Also apparent from Table 3.4.1 is that the most consistent warming across the baseline period (albeit with variable amplitudes) is centred on the 1980s for both T_{max} and T_{min}. Given the variable pattern in the 1960s, the 1980s warming obviously contributes relatively more to the total warming recorded over the baseline period for eastern stations by comparison with those further west.

3.4.1.2 Linkages to variations in other datasets

Despite the lack of data for the 1990-2000 period, it is useful nonetheless to relate broad patterns of variation in the baseline station data to temperature fluctuations in a dataset with greater spatial cover on a longer time-series. A five degree resolution CRU dataset covers a grid area which is representative of Scotland and indicates that some more recent global and hemispheric warm year anomalies can also be detected in the Scottish temperature record (Figure 3.4.3a). In terms of linking anomalies in the 1961-1990 baseline data to wider global, hemispheric and regional trends, in some respects it is unfortunate that the 1971-2000 period was not selected as the baseline period as some of the warming and wetting trends through the 1990s may have been captured. However, the reasons governing the choice of the 1961-1990 averaging period have been discussed previously (Section 2.1.1), and, as will be demonstrated in subsequent chapters, the choice of the 1961-1990 baseline period remains crucial for other aspects of the work.

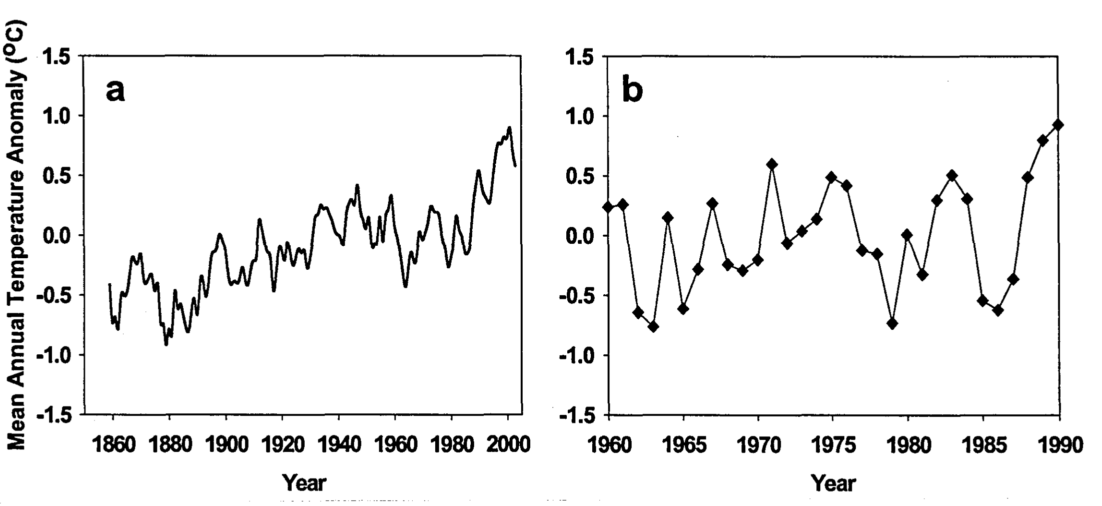


Figure 3.4.3: (a) Annual land temperature record anomalies (1857-2003) relative to 1961-1990 mean temperatures for a 5 degree grid box extending from 55° - 66° N and 0° - 5° W (Source CRU; see Jones *et al.*,1999; Jones and Moberg, 2003). These have been smoothed on a 5 year running mean. (b) Individual year anomalies for the 1961-1990 period from the same dataset

Apparent from Figure 3.4.3, is that despite regional differences in fluctuations of T_{max} and T_{min} over the 1961-1990 baseline period, the broader pattern of warming and cooling for the period broadly parallels wider regional data on longer time-series for mean annual temperature. Figure 3.4.3 also usefully illustrates that the annual temperature anomalies for the 5 degree grid box covering Scotland broadly parallel wider Northern Hemispheric temperature trends in recent years (Figures 1.1 and 1.2). The warm year anomalies associated with the 1990s and early 2000s are particularly evident in the smoothed dataset used to generate Figure 3.4.3(a).

Following the early 20th century cooling, there has been a gradual warming over the second half of the century aside from the grouping of cooler years in the early 1960s and late 1970s. This is also true for the clustering of the anomalously warm years in the 1980s through the 1990s and into the early part of the present century. These patterns are evident in the CRU dataset if decadal trends are extracted for annual mean surface temperatures:

- 1960s and 1970s decadal cooling trends are –0.02°C and –0.05°C, respectively;
- and 1980s and 1990s warming trends are +0.03°C and +0.06°C respectively.

Analyses of data from the northern hemisphere show that the increase in temperature during the 20th century is likely to be the largest of any century during the last 1000 years (Houghton, 2002), with the temperature change over the past 30-50 years unlikely to be entirely due to internal climate variability (Stouffer *et al.*, 1994; Santer *et al.*, 1996; Tett *et al.*, 1996; Tett *et al.*, 2000). Recent warming has been greater over land compared to oceans, a mean increase of 0.24°C/decade over land comparing to 0.16° C/decade over the oceans (Stott *et. al.*, 2000), with the largest increases in temperature occurring over the mid and high latitudes of the continents in the Northern Hemisphere (IPCC, 2001).

With a mean rate of increase 0.2° C/decade for near surface temperatures, this rise is unusually rapid compared to model estimates of natural variability (Hegerl *et al.*, 1996; Santer *et al.*, 1996; Tett *et al.*, 1999). However, this mean value conceals the complexity of observed climate change, and if the recent past is a guide to the future, regional climate changes will have more profound effects than the mean global warming suggests (Vaughan *et al.*, 2001). Nonetheless some caution must be applied in obtaining realistic assessments of the magnitude of possible future regional changes. As has been demonstrated here, both in the 1961-1990 baseline data and from the CRU dataset, the observed regional warming trends in the Highlands and Scotland are not of the order of magnitude cited above even for the warm decade of the 1990s.

Significantly, much of the mean global increase is associated with a stronger warming in daily minima than in maxima, leading to a reduction in the diurnal temperature range (Easterling *et al.*, 2000; Vose *et al.*, 2005). In northern and western Europe, evidence has been found of a decreasing number of frost days since the 1930s, which appears to be associated with strong increases in winter minima (Easterling *et al.*, 2000). If winter trends are extracted for the warm decade across the regional station groupings in the 1980s (Table 3.4.1), there is some support for this in a Highland context. Mean winter T_{min} increases of 0.15°C, 0.24°C, 0.22°C and 0.23°C are recorded for Region 1, 2, 3 and 4 station groupings respectively.

This temperature rise over the last 50 years is broadly accounted for by anthropogenic forcing due to the increases in greenhouse gases, together with some cooling caused by sulphate particles from increased sulphur dioxide emissions, and some small natural forcing from volcanoes and changes in solar radiation (Royal Society, 2001). These recent increases in temperature are also highly unusual in the context of reconstructions of the past 1000 years from paleodata (Mann *et al.*, 1999; Crowley, 2000). However, the increase has not been steady since the turn of the century, rather the temperature increased abruptly between about 1920 and 1945 and again from 1975 to the present, although with a contrasting pattern between hemispheres (Figure 1.1) (Jones *et al.*, 1999; Zwiers and Weaver, 2000).

While the magnitude of warming over Scotland is not of the order recorded at wider global and hemispheric scales. Nonetheless ranking the CRU data for Scotland according to average surface temperature highlights an unusual grouping of warm years in the last two decades of the 20th century, particularly for the 1990s and early 21st century. Notably six of the ten warmest years in the CRU data record have occurred since 1990 (Table 3.4.2 below), a pattern again in broad accordance with global and hemispheric trends. It is also perhaps noteworthy that five of the ten warmest years have occurred in the recording period since 1997.

Table 3.4.2: The ten hottest years from the ranked 1857 – 2003 record (Source: CRU/Scottish Executive)

Year	**Average Surface Temperature (°C)**	**Difference From 1961-1990 Average (°C)**
2003	9.18	1.31
1997	9.07	1.20
2002	9.00	1.13
1949	8.81	0.94
1990	8.80	0.93
1999	8.79	0.92
1857	8.79	0.92
1945	8.75	0.88
1998	8.73	1.20
1959	8.72	0.85

The above trends in the CRU dataset can also be set against a longer regional time-series where a number of quality control procedures have been applied to ensure data homogeneity over the instrumental record, the SNIFFER datasets (Purves *et al.*, 2000; Jones and Lister 2004 a, b). The three series for the Scottish mainland (SMT), Scottish Islands (SIT) and Northern Ireland (NIT) show long-term annual warming of 0.69 °C, 0.64 °C and 0.77 °C respectively over the 1861–2000 period (Jones and Lister 2004 a, b).

In the revised SNIFFER data, twenty-three of the warmest months, seasons and years (out of a total of 51 series: 17 time series and three locations) in the period 1861 to 2002 have occurred since 1988, while the last coldest month was September 1952 for NIT, with 30 of the 51 coldest extremes occurring in the 1861–99 period (Jones and Lister 2004 a, b). Also, given the known strong influence of the NAO on European temperatures, correlations of the order of 0.54 – 0.63 were obtained for the winter which were also weakly significant in spring (Jones and Lister 2004 a, b).

Given some of the patterns of variation reported here, not surprisingly given the wide range of observed quantities more recently analysed by Barnett *et al.* (2006), a complex picture of a changing Scottish climate emerged. However, they emphasise the difficulties of obtaining homogeneous datasets that can be gridded for mapping (an issue explored further in Section 7.3.1) and urge caution in the interpretation of some of the data (Barnett *et al.*, 2006). Despite this caveat, some of their key conclusions were:

- Significant temperature increases have occurred in recent decades, these at a faster rate than at any other time in the ninety-year period analysed, albeit with considerable variations in the pattern of regional fluctuations:
- Since 1961 average daily maxima have been increasing at a faster rate than average minima, or night-time temperatures. This concurs with some of the trends reported in Table 3.4.1 here:
- The rate of change of both annual and winter mean precipitation have also been faster since 1961 than at any other time in the 1914 to 2004 record. Again however, there is regional variation. Barnett *et al's* (2006) findings that heavy rainfall events have increased significantly in winter for northern and western Scotland are also in line with the earlier findings of Osborn *et al.* (2000) and the more recent work by Fowler and Kilsby (2003a, b).

In terms of the winter 1961-1990 baseline series used here, it is reasonable to assume that any prospective NAO influences are likely to be most readily detectable in winter station records around the west coast. Therefore the T_{max} and T_{min} anomaly values were summed and averaged (T_{mean}) for the Region 1 stations described in Section 3.4.1 and smoothed using a five year running mean. These are plotted alongside the corresponding smoothed winter (DJF) NAO Index (NAOI) values on the same running mean (Figure 3.4.4). These are widely used NAOI values obtained via the CRU website and are described further in Jones *et al.* (1997).

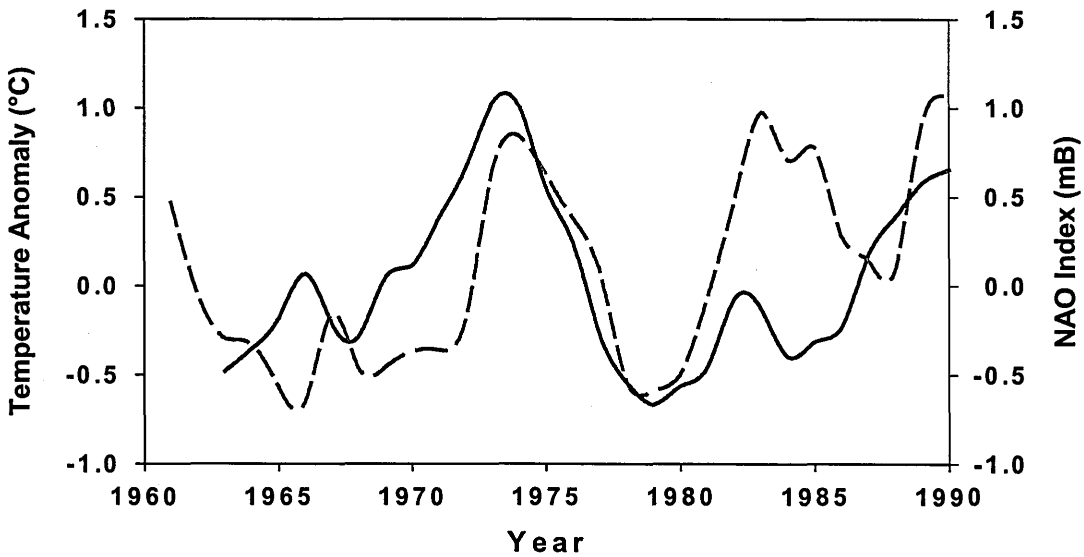


Figure 3.4.4: Winter (DJF) Region 1 1961-1990 mean temperature anomalies (blue line) compared with the equivalent winter (DJF) NAO Index series (red dashed line). Each series has been smoothed on a 5 year running mean.

As would be expected for this region, Figure 3.4.4 illustrates some broad linkages, most notably for the period centred on the 1970s. The late 70s cooling so apparent in Figures 3.2 and 3.3 is also evident and is linked in the shift of the winter NAO to a more negative phase. However, any NAO linkages associated with the mid-1980s cooling are not so obvious. In some respects this is surprising since the more progressive westerly circulation NW Europe has experienced in recent decades can be linked to the highly positive phase the NAO has exhibited since 1980 (Hurrell and Van Loon, 1997; Hurrell, 1998; Paeth *et al.*, 1999; Mayes, 2000; Marshall *et al.*, 2001).

This apparent disjunct between the regional temperature anomalies and the NAOI during the early to mid-1980s is puzzling. Not least since for all four of the regionally averaged temperature datasets across the baseline period, the largest anomalies by a considerable margin for both winter T_{max} and T_{min} recorded for 1989 appear to be associated with the unusually high NAOI winter values that year (Hurrell, 1995;1998).

Despite the above observations, in broad terms the patterns of variability are similar to those recorded by Jones and Lister (2004 a, b) in the longer records. However, correlations obtained for the Region 1 dataset here are lower for winter ($r = 0.493$; $p = 0.08$) by comparison with the reported values from Jones and Lister noted above. While for the Region 4 winter T_{mean} dataset, correlations are lower again ($r = 0.397$, $p < 0.05$). However, as Jones and Lister were working on 100+ year time-series and using station data sets from a wider geographical area, some differences could be reasonably expected.

As was previously noted above, while it is unfortunate that the wider baseline station records do not extend into the 1990s, averaging annual T_{max} and T_{min} to produce an annual mean for the four upland station records used in subsequent work (see section 5.2, Chapter 5) followed by a data ranking exercise is illuminating. While these are restricted datasets in terms of available record length (see Section 5.2 for wider discussion). For some of the stations there is some correspondence between warm years in the upland record and the CRU data tabled above, e.g.;

- 1997 was the warmest year in Cairngorm chairlift and Cairngorm siesaws record and the 3rd and 7th warmest year in the Cairnwell and Aonach Mor records respectively. These are coincident with 1997 as the 2nd warmest year in the CRU record (Table 3.4.1);

- 1990 was the 2nd warmest year in the Cairngorm chairlift record (no data available for other stations) and is coincident with the 5th warmest year in the CRU record;
- 1999 was the 2nd, 3rd and 4th warmest year at Cairnwell, Aonach Mor and Cairngorm siesaws respectively (no data for Cairngorm chairlift). These are coincident with 1999 as the 6th warmest year on the CRU record;
- 1998 was the 4th warmest year at Cairnwell, 5th warmest at Cairngorm chairlift and 6th warmest in both the Cairngorm Siesaws and Aonach Mor station data. These are coincident with 1998 being the 8th warmest in the CRU record.

Unfortunately, for most of the Highlands upland station records remain limited due to the environmental and practical difficulties of collecting reliable data. While the above rankings have no statistical rigour, it would be interesting to explore further why temperature trends at upland sites in the east are in closer accordance with wider regional trends than western sites such as Aonach Mor.

Similarly, it would have been interesting to compare the trends more closely with those of Johnston (1999) for the paired Balquihidder stations, the Lower Monachyle AWS (310m) and the Kirkton High AWS (670m). The time-series of annual temperatures showed a mean decrease of 0.1°C at the Monachyle while a rise of 0.2°C occurred at the Kirkton station – most of this in the period 1987-1990 (Johnson, 1999). The results which potentially had the most impact on these environments were the rise in temperatures during the winter months and the temporary, but large decrease in summer temperatures on the ridge-top (Johnson, 1999). However, the upland station records used later in further aspects of this work are not available on such usefully concordant time series which overlap to the same extent at appropriate locations.

Despite this there are some indications of warming at CairnGorm if trends for winter T_{max} and T_{min} are extracted from the two station records available there (Chapter 5; Sections 5.1 and 5.2);

- mean winter T_{max} and T_{min} rose by +0.08°C and +0.06°C respectively in the seventeen year (1982-1998) Cairngorm chairlift record;
- while mean winter T_{max} and T_{min} rose by +0.21°C and +0.23°C respectively in the seven year (1993-1999) Cairngorm Siesaws record.

Therefore there is some support (albeit from a short record) that the wider warming recorded through the 1990s has extended to higher elevation sites. If winter trends are extracted from the 1993-1998 segment of the chairlift station record, these are of a similar magnitude to those recorded in the 1993-1999 Siesaws record, +0.19°C and +0.21°C for T_{max} and T_{min} respectively. On a wider analysis, other workers have reported a reconstructed spatio-temporal pattern of warming for European mountain regions similar to that in the lowlands, but with an increase in the diurnal temperature range in the mountains in addition to the overall warming (Kettle and Thompson, 2004).

Despite much of the preceding commentary on the relative worth of one baseline period over another, an important cautionary note is required in relation to the use of any late 19th century baseline to model 20th century climate. The mid-19th century warm period evident in the Central England Temperature Series (CET) and elsewhere has received relatively little attention (Butler *et al.*, 2005). Consequently, the use of an end of 19th century baseline incorporating the 1970s cooling, as is the case in this work and in the wider examples cited above, will tend to exaggerate any subsequent warming in the 20th century.

3.4.2 Variability In Precipitation

Procedures here followed those described for Precip in Section 2.3. Again, while the exercise was conducted for a full seasonal range of Precip totals, only mean annual anomaly plots are presented here for the same reasons as detailed in 3.4.1 above. Similarly, no attempt was made to statistically infill missing years in individual station records for the same reasons as detailed in Section 3.4.1. Since a greater number of Precip stations were used by comparison with temperature, the station groupings used to generate the anomaly plots are summarised in Table 3.4.2.1 below;

Table 3.4.2.1: Precip station groupings used to generate baseline annual anomaly plots west-east

Western (Region 1) Station Name and ID	**Central (Region 2) Station Name and ID**	**Eastern (Region 3) Station Name and ID**
11 Allt Leachdach	14 Glen Einich	1 Lumphanan
12 Alltbeithe	15 Glenfeshie Lodge	2 Balnakeilly
13 Fort William	16 Loch a' Chrathaich	3 Barney
17 Loch Eilde Mhor	18 Loch Pattack	4 Clintlaw
19 Lochan na Sgud	20 Lon Mor	5 Glen Lee
21 Rhum, Coire Dubh	26 Sronlairig Lodge	6 Glenhead
22 Rhum, Harris	27 Tromie Dam	7 Longdrum
23 Rhum, Kilmory	28 Allt Dubhaig	8 Newton
24 Rhum, Slugan Burn	29 Allt Girnaig	9 The Linns
25 Skye, Alltdearg House	30 Allt na Bogair	10 Westerton
42 Am Meallan	31 Ben Vorlich	
43 Carn Anthony	32 Bruar Intake	
44 Coulin Pass	33 Capel Hill North	
45 Gleann Fhiodhaig	34 Capel Hill West	
46 Glenshiel Forest	35 Chapel Burn	
47 Grudie Bridge	36 Frandy Burn	
48 Heights of Kinlochewe	37 Loch Benally South	
49 Loch Fannich South	38 Loch Ordie	
51 Loch Gleann na Muice	39 Lochan Oissineach Mhor	
54 Strone Nea	40 Slateford Burn	
55 Tornapress	41 Sronphadruig Lodge	
	50 Loch Glass	
	52 Luibachlaggan	
	53 Strathrannoch	

Again, despite the rather provisional grouping of stations, the underpinning principle governing selection (representation of the west-east precipitation gradient) was the same as for the temperature stations. Similarly, the station groupings for Precip can also be linked to the CI as below;

- Western stations (Region 1: anomaly plots Figure 3.3a) are coincident with the hyper-oceanic/oceanic west and relate to CI scores of 2-6. For this station group (n = 21), latitude and longitude ranged from 56.91° - 57.81° and -6.38° – -5.02° respectively, with an elevation range of 15m – 366m.
- Central stations (Region 2; anomaly plots Figure 3.3b) are coincident with the drier inland zone with CI scores of 6-10. For this station group (n = 24), latitude and longitude ranged from 56.21° – 57.76° and -4.77° - -3.47° respectively, with an elevation range of 160m – 536m.
- Eastern stations (Region 3; anomaly plots Figure 3.3c) are coincident with the same CI zone, but with a more eastern bias. For this station group (n = 10), latitudinal and longitudinal ranges are 56.70° – 57.12° and –3.22° - -2.70° respectively, with an elevation range of 145m – 471m.

Precip stations were only grouped into three provisional regions as the spatial cover provided by stations selected for the baseline did not extend as far to the east as some of the baseline temperature stations (Figure 3.4.1). As a result, baseline station cover for the Aberdeenshire and Morayshire lowlands is poor (Figure 3.4.1). Despite the lack of station cover in the eastern lowlands, the available station records detailed in Table 3.4.2.1 were used to generate regional anomaly plots for the baseline period (Figure 3.4.5).

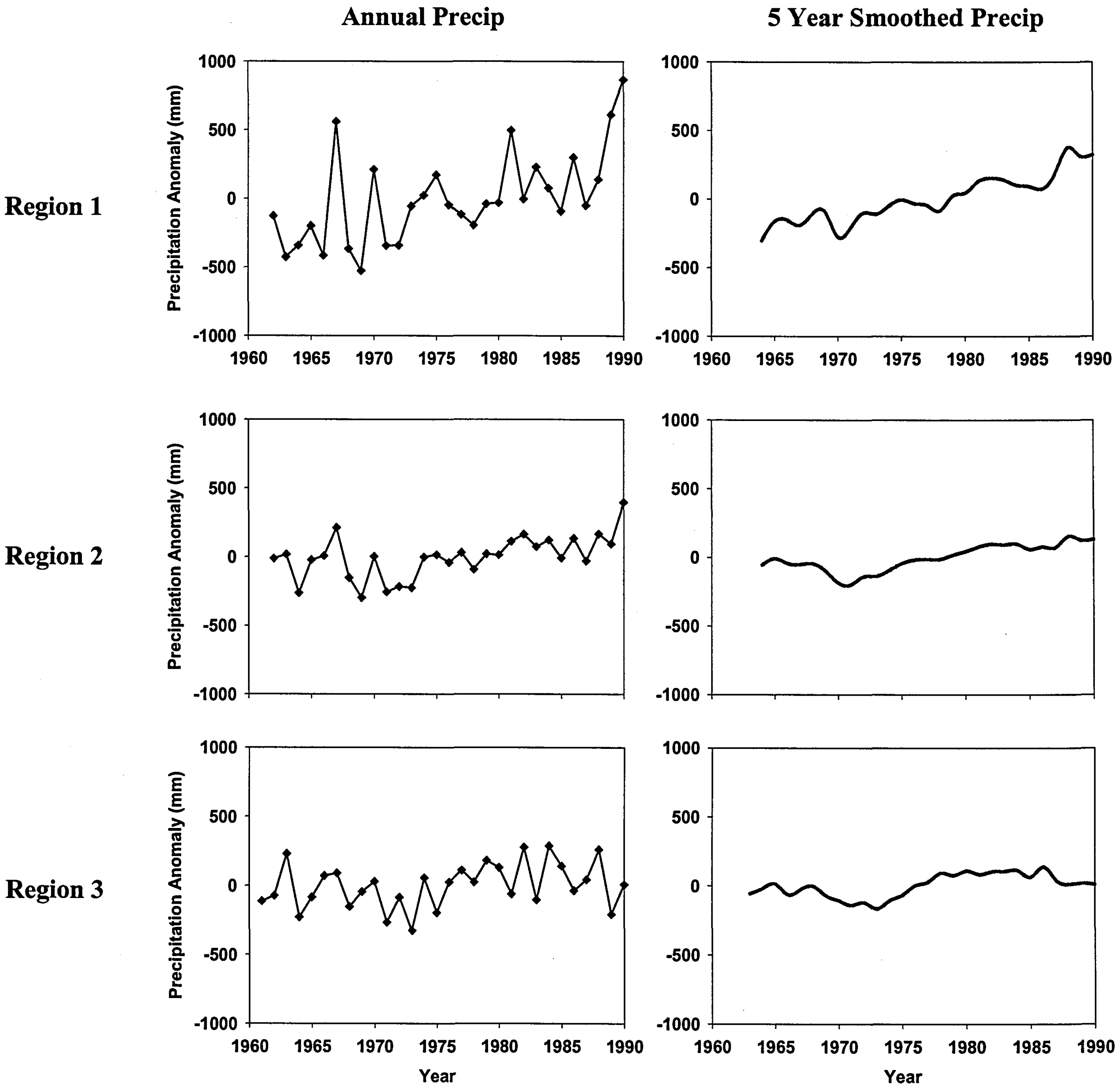


Figure 3.4.5: Regionally averaged annual Precip anomalies 1961-1990 for Region 1, 2, and 3 stations for annual anomalies and with a 5-year smoothing applied

Evident from the Precip anomaly plots for all of the regionally averaged station data (Figure 3.4.5) is the year to year and decade to decade variability in annual Precip totals, with the plots also usefully illustrating the marked precipitation gradient west – east across the region. While patterns of variability are broadly similar, the amplitude of variation for Region 1 stations is considerably greater compared to those recorded for stations in the more sheltered inland and eastern locations associated with Regions 1 and 2 (Figure 3.4.5). The five year smoothed plots indicate an upward trend in Precip from the early to mid-1970s across the Highlands, but with a more variable pattern by region through the 1980s. However, the increase in the mid-1980s is considerably more marked for Region 1 stations when compared with Regions 2 and 3. This is likely to be linked to the recent highly positive phase of the NAOI (Section 3.4.2.1).

As for temperature, these trends in the regional 1961-1990 baseline data can be related to fluctuations in a dataset with greater spatial cover and on a longer time-series (Figure 3.4.6) (Jones and Conway, 1997; Alexander and Jones, 2001).

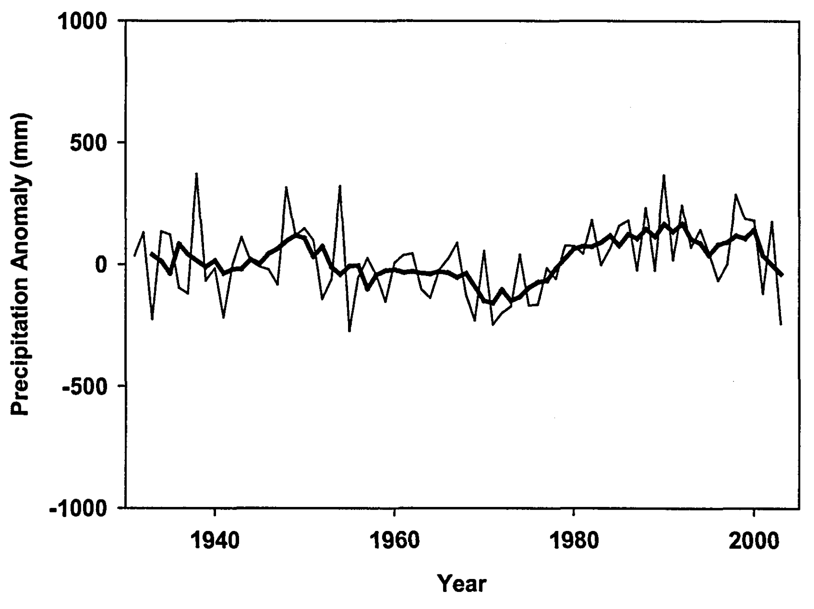


Figure 3.4.6: Annual 1931 – 2003 precipitation record anomalies (thinner line) relative to 1961-1990 mean annual precipitation for 21 Scottish sites (Raw data source CRU; see Jones *et al.*,1997). The 1961-1990 baseline was calculated and anomalies derived. The thicker line plot is for anomalies smoothed on a 5 year running mean.

Apparent from Figure 3.4.6, particularly when the smoothing is applied, is that average annual precipitation in the 1980s and 1990s was higher than in each of the previous 5 decades, particularly the 1970s, which contained several years with below average rainfall.

3.4.2.1 Linkages to precipitation variability in work elsewhere and the influence of the NAO

As was discussed in the context of temperature above, it is unfortunate that the 1971-2000 averaging period was not selected as the baseline period, at least in terms of analysing the datasets for changes detected in studies elsewhere through the late 1980s, and in particular throughout the 1990s. Changes in seasonal precipitation intensity over this period associated with changes to the NAO have been greatest in the parts of the UK which experience significant orographic rainfall (Osborn *et al.*, 2000).

It would appear therefore that the Scottish precipitation increases in the autumn, winter and spring seasons of the 1980s and through the early 1990s considered by Smith (1995) to be the largest sustained anomaly in the record, would appear to be congruent with the recent highly positive phase of the NAOI. Dawson *et al.* (2001) also considered that the recent sustained increase in the NAOI broadly reflects a pattern of increased storminess across Scotland, particularly around the coast when the Icelandic Low is well developed.

For a fuller review of recent trends in storm track changes for the North Atlantic sector, the influence of the NAO and possible future changes associated with wider global warming, the interested reader is directed to Annex Paper 3., Sections 2.2.1, 2.2.2 and 3.3. This paper also contains a section linking regional gale day frequency (as a proxy for westerly dominated winters) and the winter NAOI at a monthly resolution.

More active westerlies associated with the recent positive phase of the winter NAO have characterised much of the 1980s and 1990s across the north-eastern Atlantic sector and northwest Europe (Hurrell and Van Loon, 1997; Hurrell, 1998; Paeth *et al.*, 1999; Mayes, 2000; Marshall *et al.*, 2001). These have been coincident with increased regional gradients of rainfall and associated changes in the distribution of daily precipitation amounts in the UK over the period 1961-2000, on average becoming more intense in winter and less intense in summer (Osborn and Hulme, 2000).

These precipitation changes have been reflected in changes to flow rates in more than half of Scotland's largest rivers, particularly those in the west (Werrity *et al.*, 2002). Associated with these changes in mean values other regional frequency analysis estimates indicate that the magnitude of extreme rainfall has increased two-fold over parts of the UK since the 1960s and particularly in the 1990s (Fowler and Kilsby, 2003[a, b]). In their analysis the largest changes are observed in autumn and spring, with considerable change evident in autumn and winter extremes. In Scotland the 50 year event in 1961-90 has become an 8-year, 11-year, and 25-year event in the East, South and North Scotland pooling regions respectively during the 1990s (Fowler and Kilsby, 2003[a]).

Given the dominant influence the NAO has on many aspects of the NW European climate, some linkages with 1961-1990 Precip variations should be expected, especially in the west (Figure 3.4.6). Correlations for the Region 1 station anomalies with the corresponding winter NAOI are reasonable ($r = 0.522$, $p < 0.005$) if the unadjusted yearly values for both are used. However, the correlation improves greatly if the five year smoothed running mean values (basis of Figure 3.4.5) are used ($r = 0.770$, $p < 0.001$).

Although not charted, some NAOI linkages to winter Precip variability can also be detected for Region 2 stations. If unadjusted yearly values are used, the correlation is low and the significance value borderline ($r = 0.484$, $p = 0.08$). However, utilising the five year smoothed values again improves both the correlation and statistical significance ($r = 0.691$, $p < 0.001$). These lower correlation values for this station grouping are not that surprising. These are likely to be reflecting the effectiveness of the west coast mountains as a north-south orographic barrier to tracking westerlies in more NAO-positive winters. Repeating the correlation exercise for the Region 3 stations emphasises this point. Unadjusted yearly values return low and statistically insignificant correlation coefficients ($r = 0.172$, $p = 0.371$). While for these stations, the five year smoothed values return low correlation coefficients, albeit with statistical significance ($r = 0.392$, $p < 0.05$).

The above findings make sense when interpreted in the context of the regional climate of the Highlands. In addition, some recent modelling studies have demonstrated the important role played by topographic forcing in modulating the large-scale NAO variability over Europe, with results suggesting that the interaction of large-scale

NAO type circulations with topography is generally more important for precipitation than for temperature (Bojario *et al.*, 2005).

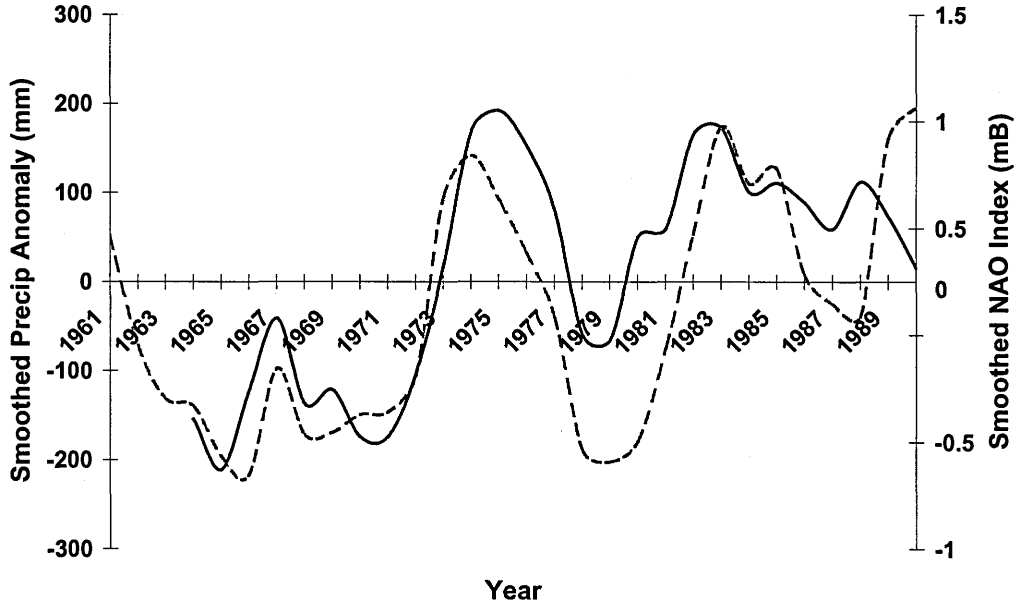


Figure 3.4.7: Winter (DJF) Region 1 station 1961-1990 mean precipitation anomalies (blue line) compared with the equivalent winter (DJF) NAO Index series (red dashed line). Each series has been smoothed on a 5 year running mean.

Figure 3.4.7 provides a useful illustration of the pattern of variation between the datasets. The close fits around ~1975 and ~1983 (above are smoothed datasets) are of interest. When the anomaly data for both the Region 1 and Region 2 stations are ranked, 1975 and 1983 are the 1st and 2nd wettest years respectively in the 1961-1990 record. The NAOI winter value in 1975 was unusually high (~2.5), with winter NAOI values >2 generally being linked to exceptionally mild winters and strong westerlies (Jones and Lister, 2004,[a, b]).

3.4.2.1 Detecting precipitation variability in extended time series

Despite the opening comments to this section on the lack of analysis of data sets extending through the 1990s, a small number of focused and closely supervised one month undergraduate projects examined aspects of variability in precipitation across the region over time-series extending further into the 1990s for selected station locations. While not, *sensu stricto*, part of the core objective of this project, these nonetheless detected between-period Precip changes which may be significant in the context of regional change, and which may flag trends in future seasonal changes associated with wider climatic change in the Highlands. Findings are summarised by project below:

1. Precipitation change in the North of Scotland (Lasserre and Coll, 2001)

(i) **Objective**: To determine if detectable changes in seasonal precipitation gradients had occurred for a northern Highland (Caithness and Sutherland) study region over a 1985-1999 averaging period compared to 1970-1984.

(ii) **Method**: Stations with a more or less intact record, 1970-1999 (four in Sutherland and four in Caithness) representing proximal west-east transects across the counties were selected. Station data were quality-controlled and averaged to standard climatological seasons as per Section 2.2. Between-station and between-period differences were analysed for each season.

(iii) **Results**: Station data were typified by high inter-annual variability for all seasons (although spring datasets were less noisy). Consequently the skew of data dictated a non-parametric test, applied Mann-Whitney tests indicated ($p < 0.05$);

- no statistically significant Precip increases 1985-1999 compared to 1970-1984 for the Sutherland (more westerly) stations;
- spring Precip increases 1985-1999 versus 1970-1984 were statistically significant. These spring increases also applied to Caithness (more

easterly) stations and extended into summer for the two most easterly Caithness stations.

2. Detecting changes in upland precipitation in the Scottish Highlands associated with climate change (Cazenave, Marino and Coll, 2001)

(i) **Objective**: To determine if detectable changes in seasonal total Precip had occurred for selected upland stations across the region for between period testing 1968-1998

(ii) **Method**: Upland station records (320m-475m) with more or less intact data 1968-1998 for ten locations west to east across the region were selected. Station data were quality controlled as described previously, the period tested was dictated by the quality of the station records. Most stations were missing substantial chunks of data for the years 1994 and 1995. In addition, a number of station records were incomplete for some of the years 1968-1994. Consequently, tests were conducted on seasonal and annual mean precipitation data for the two thirteen year periods 1968-1980 and 1981-1993 for eight stations with complete data for this period. It was reasoned that these two periods may capture possible seasonal variations as a result of higher winter NAOI values over the later testing period.

(iii) **Results:** As for Project 1 above, station data were typified by high inter-annual variability for all seasons. Therefore a two sample Mann-Whitney U-Test was again applied, since;

- by comparing median values it remains resistant to outliers and other extreme values;
- it can be applied regardless of sample size.
- Results by season are summarised in Table 3.4.2.2 below;

Table 3.4.2.2: Results of Mann-Whitney tests by station 1968-1980 versus 1981-1993; significant versus non-significant difference ($p < 0.05$), total Precip by season for stations tested west – east. Y denotes a statistically significant between-period change, N denotes no statistically significant change.

Station	Annual	Spring	Summer	Autumn	Winter
Coire Dubh (Rhum)	Y	Y	N	Y	N
Harris (Rhum)	Y	Y	N	Y	N
Loch Eilde Mhor	Y	Y	N	Y	Y
Loch a Chrathaich	Y	N	N	Y	Y
Sronlairig Lodge	Y	N	N	Y	Y
Loch Pattack	Y	Y	N	Y	Y
Lochan Oissineach Mhor	Y	N	N	Y	N
Faskally	Y	Y	N	N	N
Glen Lee	N	N	N	Y	N

Apparent from Table 3.4.2.2 is that mean annual Precip increases for all stations (with the exception of Glen Lee) were statistically significant for the later averaging period. For the majority of the stations, most of these increases were largely attributable to changes in the autumn and spring, with no statistically significant changes in summer, while winter changes were only detected in half of the station totals. It would appear that trends for these upland locations and for some of the Caithness stations in these seasons are paralleling some of the findings of Fowler and Kilsby (2003a, b) and some of the earlier findings of Jones and Conway (1997).

3.5 Recent global trends and modelled regional futures

Recent temperature trends and the grouping of warm years in recent decades are unusual in both instrumental and longer-term proxy reconstructed temperature records (Figure 1.2). Despite the various problems with climate models and widely varying outputs on the magnitude of future changes, there is agreement that the observed warming trends are set to continue throughout the 21st century. Encouragingly, much of what climate model studies show could happen to weather and climate extremes in the future with increased GHGs is what would intuitively be expected from our understanding of how the climate system works (Easterling *et al.*, 2000).

For example, an increase of GHGs produces increased surface heating with warmer surface temperatures, more evaporation, an increase in the ability of the atmosphere to hold more moisture, and thus an increase in atmospheric moisture content with enhanced precipitation rates (Trenberth *et al.*, 1999). In addition, a number of changes in future weather and climate extremes from climate models have already been seen in observations in various parts of the world, e.g.;

- decreased diurnal temperature range;
- warmer mean temperatures associated with increased extreme warm days and decreased extreme cold days;
- increased intensity of rainfall events (Easterling *et al.*, 2000).

Aside from changes in mean seasonal values, climate models predict increasingly frequent extreme weather events. For example, recent modelling results suggest that future heatwaves will increase in the second half of the century (Meehl and Tebaldi,

2004) as anthropogenic interference in the climate system becomes increasingly noticeable (Stott *et al.*, 2004).

Climate model integrations also predict increases in both the frequency and intensity of heavy rainfall in the high latitudes of the Northern Hemisphere under enhanced greenhouse conditions (McGuffie *et al.*, 1999; Jones and Reid, 2001; Palmer and Raisanen, 2002). While outputs from the more recently developed RCMs indicate substantial changes to seasonal rainfall regimes and an increase in extreme rainfall events (Christensen and Christensen, 2002; Frei, *et al.*, 2003; Huntingford *et al.*, 2003), with substantial changes in future frequencies projected for the NW European domain (Kundzewickz, *et al.* 2006). Other recent modelling work using outputs from the HadRM3 RCM indicates a 30% increase in the return period magnitude of a 10 day rainfall event in Scotland, this constituting the greatest relative change across the UK (Ekstrom *et al.*, 2005).

While these projections from climate models are consistent with recent increases in rainfall intensity seen in the UK (Osborn *et al.*, 2000; Fowler and Kilsby, 2003a,b); Europe (Brunetti *et al.*, 2000; Frei and Schar, 2001) and worldwide (Iwashima and Yamamoto, 1993; Karl and Knight, 1998; Zhai *et al.*, 1999). It is not yet possible to relate one to the other, as cause and effect. As some of the brief examinations of variability in the datasets above has demonstrated, climatic variables here are particularly noisy on various timescales. Consequently, trend interpretations and cause and effect assignations must be treated with considerable caution.

Moreover for a region such as the Highlands there are very real concerns surrounding the performance of climate models (Chapter 4 Introduction) . However, this is not a problem exclusive to this region, regional climate prediction is a problem characterised by inherent uncertainty (Mitchell and Hulme, 1999; Hulme *et al.*, 2002 Jenkins and Lowe, 2003; Willows and Connell, 2003) (Section 1.4).

Consequently, subsequent sections of this work will focus on how and if improved local projections of future climate might be obtained for this region. Following from the various modelling experiments conducted and an assessment of the local implications (Chapters 4-6) will be some considerations on how and if wider uncertainty can be integrated into regional response strategies given the extent of current gaps in knowledge (Chapter 8).

4. Evaluating the UKCIP02 HadRM3 Simulated 1961-1990 Outputs

Synopsis

The major aims of this Chapter are:

- To use the station temperature and precipitation baseline data described in previous sections to assess the performance of the HadRM3 RCM for selected locations across the region.
- Following from this assessment, to apply, test and refine temperature lapse rate models (LRMs) both to improve the validity of the inter-comparison and for application in subsequent sections.
- To then examine methods of orographic adjustment to precipitation data for application in subsequent work.

Experimental Background: Climate Model Limitations In Topographically Diverse Regions

Even with the evolution of ever more complex and sophisticated GCMs, issues remain concerning their robustness (Chase *et al.*, 2004) and their reproduction of the detail of regional climates remains rather poor (Zorita and von Storch, 1999; Gonzalez-Rouco *et al.*, 2000; Jones & Reid, 2001). For the North Atlantic region, GCMs generate widely varying results and have problems in simulating the present climate of the region (Saelthun & Barkved, 2003), while many model-to-model differences in predictions lies in the construction of the models.

This is due, in part, to a limited parameterisation of many specific climate processes since climate models have to represent all the elements of the climate system, for example; atmosphere, ocean, land surface, cryosphere, and biogeochemical cycles. The largest part of climate change arises not from the direct effect of increasing greenhouse gases but from the interaction between different components of the climate system, which give rise to a large number of positive and negative feedbacks (Jenkins and Lowe, 2003). For example, in the present climate high (ice crystal) clouds act to warm climate whereas low (water droplet) clouds act to cool climate (Jenkins and Lowe, 2003). If warming changes the characteristics of these clouds then this would have a considerable and contrasting (i.e. warming or cooling) feedback on the eventual climate change. Consequently, these are represented in different ways in the various GCMs, resulting in different outputs (Jenkins and Lowe, 2003).

Thus, when nine current GCMs are inter-compared at the UK regional scale, there are substantial differences in simulated changes to seasonal precipitation for example (Jenkins and Lowe, 2003) (Figure 1.8). As a result, more recent approaches have aimed at reducing some of the uncertainties (Murphy *et al.*, 2004) and increasing confidence in the projections of future climate (Johns *et al.*, 2003; Kerr, 2004; Stocker, 2004).

With RCMs the horizontal resolution is increased up to the mesoscale (~2 - 200km; Emanuel, 1986) over a limited area of interest allowing a much more accurate description of topography, coastlines and lakes (Christensen, 2001; Jenkins *et al.*, 2003). However, RCMs take their boundary values from the coarse grid GCMs and the effects of these coarse resolution boundary values can propagate within the RCMs, thereby perpetuating the uncertainties originating in the large scale models

(Carter *et al.*, 1999; Jenkins and Lowe, 2003; Moberg and Jones, 2003; Saelthun & Barkved, 2003; Rowell, 2004; Schiermeier, 2004b). For example, the UKCIP02 climate change scenarios were generated using the HadRM3 RCM, driven by predictions from the HadCM3 GCM. However, HadCM3's representation of some aspects of observed climate is not realistic e.g. storm tracks over north-west Europe are displaced too far south (Hulme *et al.*, 2002). This is due to a systematic error in the atmosphere only intermediate-scale component of the model-nesting (HadAM3H) where there is an incorrect high pressure bias at high latitudes for most of the year (Pope *et al.*, 2000). This affects both the pole and the Icelandic low in the Northern Hemisphere. Associated with these biases are easterly biases in the surface winds (Pope *et al.*, 2003; Christensen, 2004).

Therefore, in order to improve the coupling of the models, an intermediate, 150 km resolution, atmosphere-only GCM was used (HadAM3H) which is of intermediate scale between the coarser resolution HadCM3 and the RCM to provide the lateral boundary conditions (Gordon *et al.*, 2000; Pope *et al.*, 2000; Hulme *et al.*, 2002). This double-nesting approach improves the quality of the simulated European climate – the position of the main storm tracks, for example, are better located – and also allows the UKCIP02 scenarios to present information with greater spatial detail, 50 km rather than 250 to 300 km (Hulme *et al.*, 2002). However, in an area of complex topography such as the Highlands where the close spatial juxtaposition of sea and land areas give rise to highly localised climatic effects; it is unlikely that even such a sophisticated model will capture the detail of local climatic processes here.

Given some of these well documented problems, a key aim of this Chapter as stated above is to use the observed station data described in Sections 2.1 and 2.2 to assess

HadRM3's performance in simulating the 1961-1990 baseline for seasonal mean values of temperature and precipitation. Following from the UKCIP02 data extraction and handling procedures described in Sections 2.4.1 and 2.4.2 and the spatial referencing method described in Section 2.4.3, the observed 1961-1990 station data were used to inter-compare HadRM3's performance in simulating selected 1961-1990 annual and seasonal values of T_{max}, T_{min} and Precip across the region. Outputs from these experiments will be used to inform subsequent work sections (Chapter 5) on whether these RCM outputs and observed climatic data can be variably combined to improve local representations of future changes to temperature and precipitation.

4.1 Station Observed And HadRM3 Simulated 1961-1990 Temperature Comparisons

4.1.1 Method I Results: Island station records included and without temperature lapse-rate adjustment

Following the protocols and procedures described in Sections 2.4.1 and 2.4.2, UKCIP02 HadRM3 output data for land grid cells corresponding to the 'Highlands' (Figure 2.1; Table 2.2) were extracted to data directories. For purposes of the experiments documented here, data outputs for HadRM3's simulation of mean seasonal T_{max} and T_{min} for 'land' grid cells across the region were used. Using the method described in Section 2.4.3, station locations for the identified baseline records were interpolated to the nearest corresponding HadRM3 grid cell (Appendix Figure 2.1).

Following the procedures and reasoning described in Section 2.4.4.1, HadRM3 grid cell-simulated 1961-1990 values for mean seasonal T_{max} and T_{min} were subtracted from the corresponding observed station values and the differentials plotted (Figure 4.1).

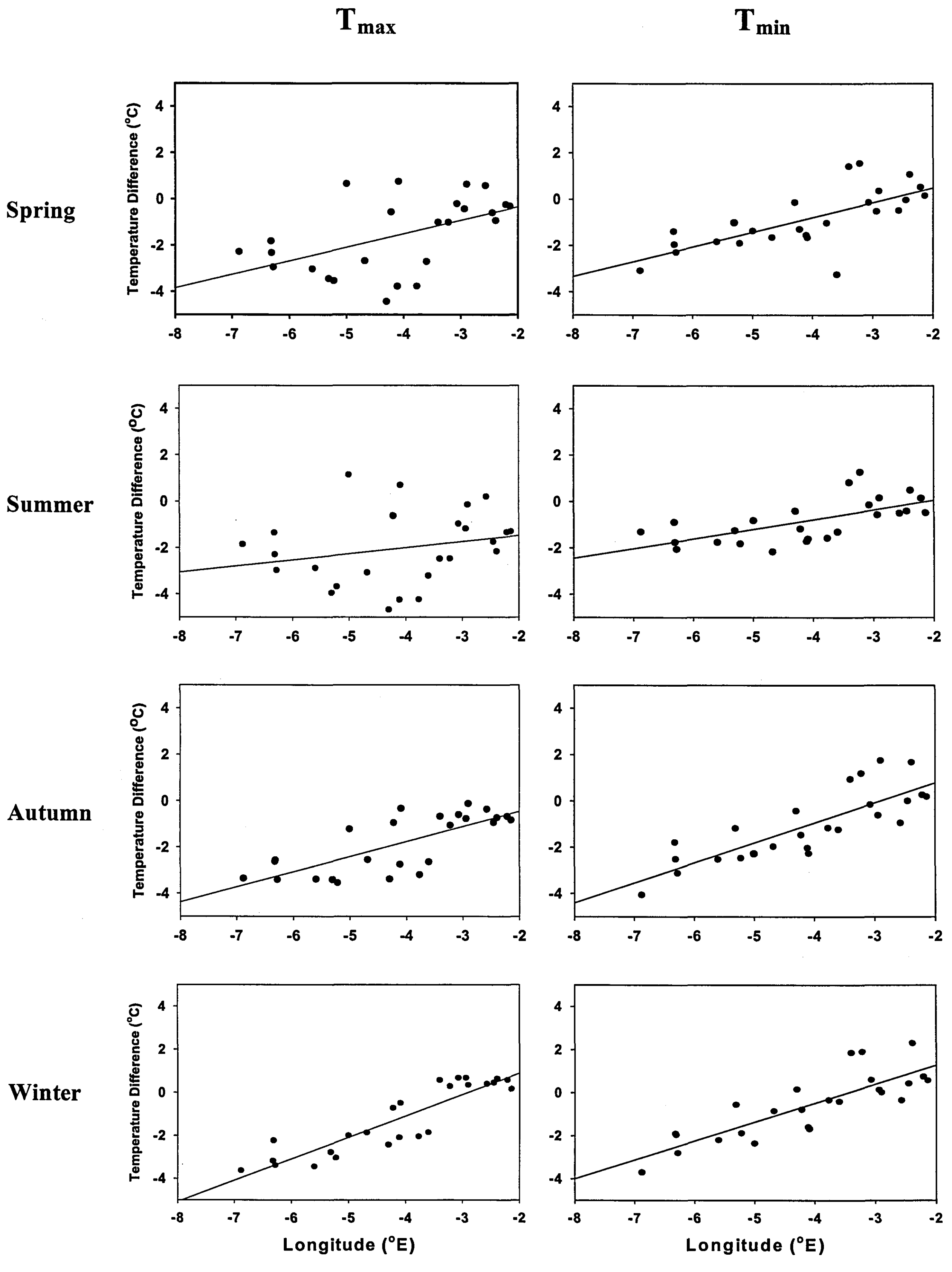


Figure 4.1: Temperature differences (°C) HadRM3 – observed station values (n = 25).

Pearson correlation coefficient (r) and probability (p) values were also calculated for each dataset between the mean seasonal values of observed and HadRM3 simulated seasonal T_{max} and T_{min}. These are recorded in Table 4.1.

Table 4.1: Pearson correlation coefficient (r) and probability values (p) - station observed and HadRM3 simulated mean seasonal T_{max} and T_{min} without lapse rate adjustment

Variable	Spring	Summer	Autumn	Winter
Mean T_{max}	r = -0.107, p = 0.612	r = 0.230, p = 0.269	r = 0.242, p = 0.244	r = -0.072, p = 0.732
Mean T_{min}	r = 0.180, p = 0.390	r = 0.382, p = 0.060	r = -0.031, p = 0.884	r = 0.142, p = 0.499

It is apparent from Figure and Table 4.1 that for all stations there is a west to east gradient of departures in seasonal maxima and minima from the corresponding HadRM3 grid cells, but none of the correlations are statistically significant (p = 0.060 – 0.884). However, HadRM3's simulation of winter maxima for example has less scatter (r = -0.072, p = 0.732) west to east than for the simulation of spring and summer maxima (r = -0.107, p = 0.612; r = 0.230, p= 0.269).

For the region as a whole, HadRM3's closest simulation of the observed baseline is for the eastern lowlands, whereas the largest departures are associated with the western lowlands and central mountainous areas. At least part of this pattern is attributable to the difference in altitude between the stations and their corresponding HadRM3 grid cell which are generally greater in western and central inland areas of the region. This can be illustrated with reference to annual T_{max} and T_{min} where temperature differentials (HadRM3 – observed station values) are plotted against the elevation difference between the HadRM3 grid cells and the stations (Figure 4.2).

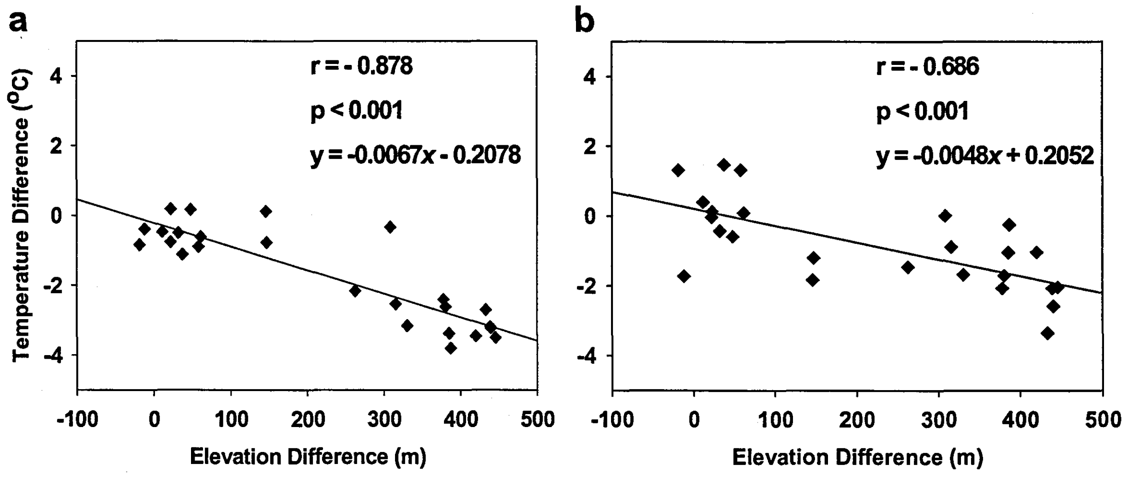


Figure 4.2: Scatter plot of annual mean (a) T_{max} and (b) T_{min} differences against elevation difference HadRM3 grids and stations

4.1.2 Method II: Island station records excluded and with temperature lapse-rate adjustment applied

Procedures here and their underpinning reasoning have been described in Section 2.4.4.2. For this analysis, HadRM3 grid cell-simulated 1961-1990 values for T_{max} and T_{min} were subtracted from the infilled and lapse-rate adjusted observed station values and the differentials plotted (Figure 4.3). This was in addition to removing the island sites from this analysis, hence why n = 20 in Figure 4.3. Pearson correlation coefficient (r) and probability (p) values were again calculated for each dataset between the mean seasonal values of observed and HadRM3 simulated seasonal T_{max} and T_{min}. These are recorded in Table 4.2.

Table 4.2: Pearson correlation coefficient (r) and probability values (p) - station observed and HadRM3 simulated mean seasonal T_{max} and T_{min} with lapse rate adjustment

Variable	Spring	Summer	Autumn	Winter
Mean T_{max}	$r = 0.786$, $p < 0.001$	$r = 0.672$, $p = 0.001$	$r = 0.921$, $p < 0.001$	$r = 0.894$, $p < 0.001$
Mean T_{min}	$r = 0.847$, $p < 0.001$	$r = 0.924$, $p = 0.001$	$r = 0.626$, $p = 0.003$	$r = 0.707$, $p < 0.001$

It is clear from Table 4.2 and Figure 4.3 that the combination of infilling the station baseline data and the lapse rate adjustment greatly improves the correlation coefficient values obtained between the HadRM3 simulated and station values, as well as returning statistically significant results. It should be noted that most of this improvement is due to the lapse rate adjustment to station baselines. As was recorded in Section 3.1, the baseline value changes associated with the infilling exercise were not that substantial.

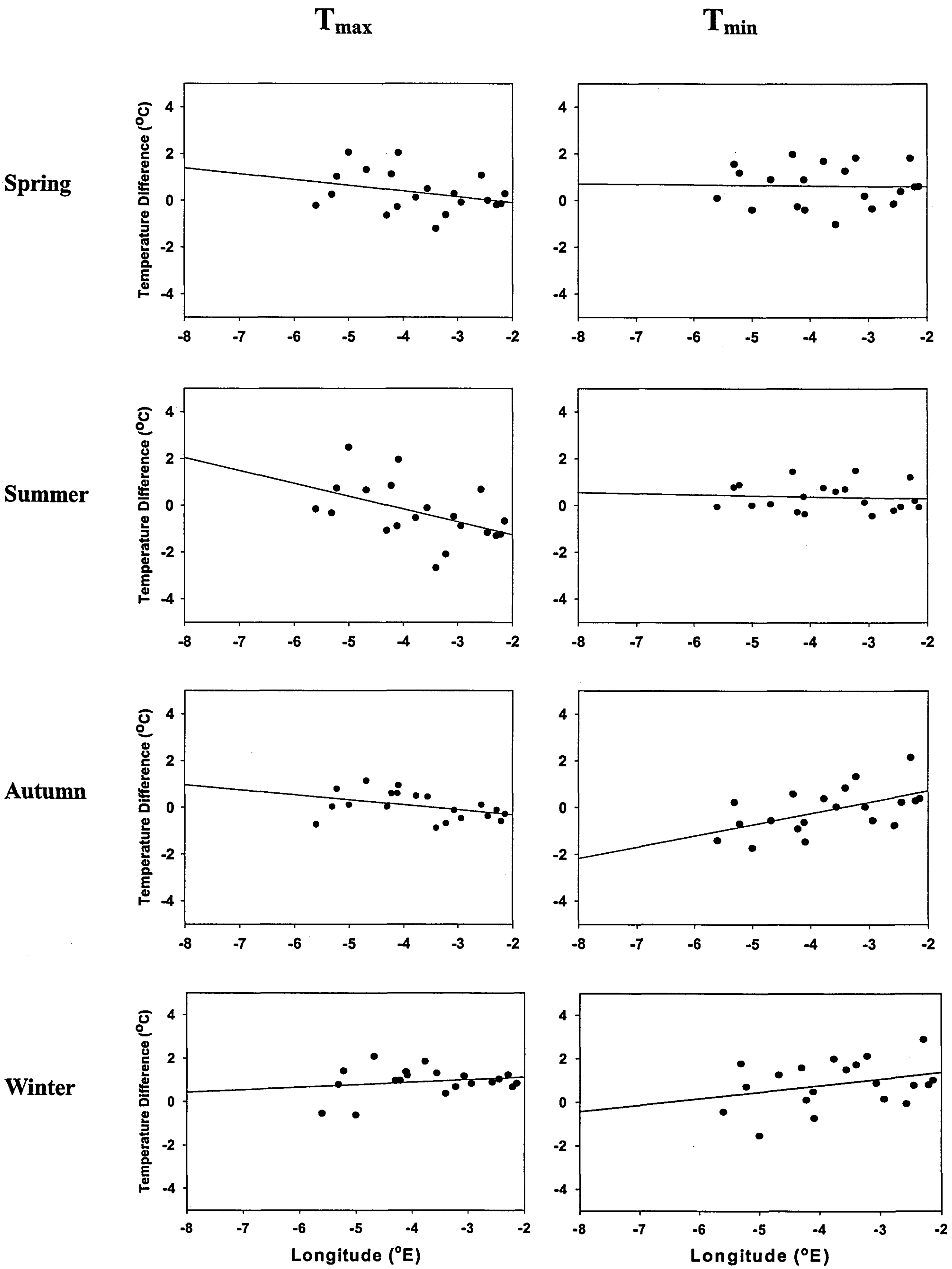


Figure 4.3: Temperature differences (°C) HadRM3 – observed station values (n = 20)

It is apparent from Table 4.2 that these improved correlation values are consistent across the full seasonal range for both T_{max} (r = 0.672 – 0.921; p = <0.001 - = 0.001) and T_{min} (r = 0.626 – 0.924 ; p = <0.001 - = 0.003). However, there is still a residual scatter of data for certain seasons, most notably for autumn T_{min} (r = 0.626, p = 0.003), summer T_{max} (r = 0.672, p <0.001) and spring T_{max} (r = 0.786, p <0.001) (Table 4.2). Whereas for other seasons, there is a much closer accordance of values and considerably less scatter, particularly for summer T_{min} (r = 0.924, p <0.001), autumn T_{max} (r = 0.921, p <0.001) and winter T_{max} (r = 0.894, p <0.001) (Table 4.2).

Despite these much closer correlation coefficient values, the results tend to suggest parameterisation problems with some of the HadRM3 grid cells. While some of the seasonal disparities for both T_{max} and T_{min} (Tables 4.3 & 4.4) may be due to the application of the lapse-rate models in adjusting the observed baseline data, it is unlikely that some of the biggest differences are due solely to the lapse-rate values applied (see also section 4.1.3). The performance of the lapse-rate models against some available upland station records also involving a refinement to the approach used here (mean lapse rate range only) is explored further in Section 5.2.

Therefore, given the widely recognised problems with climate models in topographically diverse regions, problems with the parameterisation of some of the HadRM3 grids seems a reasonable supposition. To support this point, even when a finer-scaled version of the Hadley Centre Regional climate model on a 25km x 25km grid was used to simulate climate change on some of the islands around the UK (Jenkins *et al.*, 2003), the model resolution was still unable to take account of smaller features which produce local climates, such as hills and valleys.

Table 4.3: Seasonal mean T_{max} differences HadRM3 grid cell simulated – station observed for the 1961-1990 baseline

ID Number	Station Name	HadRM3 Grid ID Number	Temperature Difference by Season (°C)			
			Spring T_{max}	Summer T_{max}	Autumn T_{max}	Winter T_{max}
1	Aberdeen	146	0.29	-0.66	-0.27	0.85
2	Balmoral	145	-0.62	-2.09	-0.68	0.69
3	Craibstone	146	-0.14	-1.23	-0.58	0.69
4	Fyvie Castle	146	-0.19	-1.29	-0.11	1.24
5	Arbroath	181	1.08	0.69	0.12	0.90
6	Dundee	181	-0.08	-0.87	-0.46	0.84
7	Montrose	146	0.00	-1.16	-0.36	1.05
8	Mylnefield	181	0.29	-0.48	-0.12	1.18
9	Fort Augustus	143	1.32	0.66	1.15	2.09
10	Inverness	126	1.12	0.84	0.60	0.98
12	Onich	160	1.03	0.74	0.80	1.42
14	Rannoch School	161	-0.64	-1.07	0.04	0.97
15	Faskally	162	0.13	-0.53	0.50	1.86
16	Fortrose	126	2.04	1.96	0.95	1.22
17	Kinlochewe	125	0.26	-0.32	0.04	0.80
18	Poolewe	124	-0.22	-0.15	-0.72	-0.54
19	Cape Wrath	108	2.06	2.49	0.12	-0.61
21	Ardtalnaig	162	-0.28	-0.89	0.61	1.38
22	Braemar	145	-1.21	-2.67	-0.87	0.38
23	Kinloss	145	0.49	-0.11	0.46	1.32

Some of the values for seasonal T_{max} differences in Table 4.3 support the interpretation offered above. Valley sites surrounded by hills where locally resolved seasonal orographic effects are most pronounced throw up some of the biggest disparities, suggesting that these localised processes are not adequately captured in the parameterisation of some HadRM3 grids. For example there is considerable over-estimation of winter and spring T_{max} in the grid cell corresponding to Onich and Fort Augustus. Similarly, there are over-estimations of winter T_{max} in other HadRM3 grid cells corresponding to other valley sites subject to orographic cooling under certain atmospheric conditions (see also Section 3.1.1), such as Ardtalnaig and Faskally.

Parameterisation problems also seem to be evident in the simulation of spring T_{max} for HadRM3 grids corresponding to certain sites. For example spring T_{max} is under-estimated in grid cells corresponding to valley sites subject to frost effects such as Rannoch School, Ardtalnaig, Braemar and Balmoral (Section 3.1.1). Similarly, the under-estimation of summer T_{max} for grid cells corresponding to Kinlochewe, Rannoch School, Faskally, Braemar and Balmoral would suggest a problem with HadRM3's ability to capture local valley heating effects in summer (see Section 4.1.3).

Local coastal effects, particularly around the North Sea coast also do not seem to be adequately captured in the parameterisation of some HadRM3 grids. For example, there is consistent over-representation of winter T_{max} at Inverness, Fortrose, Kinloss, Mylnefield, Arbroath, Dundee and Aberdeen. Taken together with the over-representation of spring T_{max} at Inverness, Fortrose, Kinloss, Mylnefield, Arbroath and Dundee, this would seem to suggest a problem with HadRM3's resolution of local seasonal effects around the North Sea coast. Differences observed at west coast sites such as Cape Wrath and Poolewe would also seem to suggest parameterisation problems around the land-sea interface in some of the HadRM3 grids. There is considerable over-estimation of spring and summer T_{max} in the HadRM3 grid cell corresponding to Cape Wrath for example.

Comparison of some of the differences for seasonal T_{min} (Table 4.4) would seem to further support some of the interpretation offered above. For example, the consistent over-estimation of winter T_{min} in HadRM3 grids corresponding to valley sites, most notably at Kinlochewe, Onich, Fort Augustus, Rannoch School, Faskally, Braemar, Balmoral and Fyvie Castle strongly suggest that HadRM3 grid parameterisations are not capturing local valley cooling effects (see Section 4.1.3).

Table 4.4: : Seasonal mean T_{min} differences HadRM3 grid cell simulated – station observed for the 1961-1990 baseline

ID Number	Station Name	HadRM3 Grid ID Number	Temperature Difference by Season (°C)			
			Spring T_{min}	Summer T_{min}	Autumn T_{min}	Winter T_{min}
1	Aberdeen	146	0.61	-0.06	0.4	1.02
2	Balmoral	145	1.84	1.51	1.35	2.14
3	Craibstone	146	0.59	0.20	0.30	0.81
4	Fyvie Castle	146	1.82	1.21	2.16	2.89
5	Arbroath	181	-0.14	-0.2	-0.75	-0.05
6	Dundee	181	-0.34	-0.43	-0.54	0.17
7	Montrose	146	0.39	-0.05	0.24	0.79
8	Mylnefield	181	0.21	0.15	0.05	0.9
9	Fort Augustus	143	0.90	0.07	-0.55	1.27
10	Inverness	126	-0.25	-0.27	-0.89	0.12
12	Onich	160	1.19	0.89	-0.68	0.72
14	Rannoch School	161	1.99	1.46	0.60	1.60
15	Faskally	162	1.70	0.77	0.40	2.00
16	Fortrose	126	-0.39	-0.35	-1.45	-0.72
17	Kinlochewe	125	1.57	0.79	0.24	1.78
18	Poolewe	124	0.11	-0.04	-1.40	-0.44
19	Cape Wrath	108	-0.39	0.02	-1.72	-1.52
21	Ardtalnaig	162	0.90	0.39	-0.62	0.50
22	Braemar	145	1.28	0.71	0.86	1.74
23	Kinloss	145	-1.01	0.61	0.05	1.51

This interpretation is further supported with reference to the considerable over-estimation of spring T_{min} in the grid cells corresponding to sites such as Kinlochewe, Onich, Rannoch School, Braemar, Balmoral and Fyvie Castle (see preceding discussion in Section 3.1.1 and Section 4.1.3). If summer T_{min} over-estimations for grid cells corresponding to, for example, Rannoch School, Braemar, Balmoral and Fyvie Castle are compared to the considerable under-estimations of summer T_{max} at the same locations. This would also tend to suggest that the HadRM3 grid parameterisations are struggling to capture local controls on climate, particularly for central and eastern valley locations.

As for T_{max}, the over and under-estimation of winter and spring T_{min} at sites such as Inverness, Fortrose, Kinloss and Aberdeen would seem to support the interpretation that there is a problem with the resolution of localised North Sea coastal effects in these seasons. While for HadRM3 grids corresponding to east coast stations such as Inverness, Fortrose, Dundee and Arbroath, these over and under-estimations are also evident for summer T_{min}, and particularly, autumn T_{min}.

By contrast, for the HadRM3 grids corresponding to Cape Wrath and Poolewe in the west, the most significant differences are in the under-estimation of both autumn and winter T_{min}. While it cannot be proven experimentally here, perhaps these seasonal differences associated with grid cells adjacent to coastal areas in the model flag a wider problem in the coupling of the ocean and atmosphere components of the driving AOCGCM (HadCM3). Although only tentative, results obtained here for both coasts suggest at least for coastal waters, that the nested HadRM3 grids driven by the boundary conditions of the intermediate-scale HadAM3H may not be capturing the seasonal disparity and lag between sea temperatures.

This could suggest a synchronisation problem with model couplings between the ocean and atmosphere slab components of the driving HadCM3 AOCGCM. If such a problem exists, it is more likely to be in the ocean and atmosphere coupling of the coarser scale HadCM3 coupled model, since for the UKCIP02 scenarios, boundary conditions are derived from the global atmosphere model HadAM3H (Pope *et al.*,

2000) which is of intermediate scale between the coarser resolution HadCM3 and the RCM (Fowler *et al.*, 2005).

However, there are also known cold bias problems with HadAM3H in the Northern Hemisphere during the winter, particularly with HadAM3H's representation of the surface field (Pope *et al.*, 2000). With the development of the new HadAM3P (Jones *et al.*, 2005), these sort of problems may be resolved as it provides a relatively realistic climate over the European area which represents a considerable improvement over the parent coupled model (HadCM3) (Rowell, 2005).

Thus HadRM3's under-estimation of autumn and winter T_{min} may be associated with the driving AOCGCM (HadCM3) not adequately capturing the lagged thermal signature associated with Gulf Stream mixing for these seasons at an adequately resolved local scale. Similarly, the over and under-estimations in other seasons for HadRM3 grids corresponding to east coast stations may be reflecting a wider problem in how the driving HadCM3 captures the seasonal warming and cooling of the North Sea basin. The performance of RCMs over a north western European domain is explored further in Section 7.5.

4.1.3 Using 1961-1990 UKMO station data at monthly resolution for comparison with HadRM3 simulations

Given preceding discussion on some of the apparent problems associated with HadRM3 parameterisations capturing localised valley effects, this short section uses existing UKMO station data for selected locations to examine differences at a monthly rather than seasonal resolution. While similar such data are available for mainland stations at Kinlochewe, Kinloss, Ardtalnaig and Dunstaffnage (see:

http://www.met-office.gov.uk/climate/uk/averages/19611990/index.html). These were not used as the sizeable elevation disparities, station – HadRM3 grid would have required a substantial temperature adjustment via the lapse rate models. While such data is available for Craibstone and the station elevation (102 m) is close to the corresponding HadRM3 grid (ID146) at 113.38 m, the data was not used as seasonally-resolved disparities here have already been recorded in Section 4.1.2 (Tables 4.3 and 4.4).

Therefore, to minimise any inter-comparison bias associated with lapse-rate correction methods, two UKMO stations with available 1961-1990 data and with elevations close to the corresponding HadRM grid were selected. These are;

- Braemar on Deeside at 339 m is ~19 metres higher than Had Grid 145 (320.73 m);
- Kinbrace in Surtherland at 103 m is ~34 metres higher than Had Grid 110 (68.58 m).

In addition, and given preceding comments on the HadRM3 grid parameterisations in relation to localised frost hollow effects in winter and spring, both stations are subject to these effects. Mean monthly observed T_{max} and T_{min} values were extracted from the station record and plotted alongside the corresponding HadRM3 grid-simulated values (Figure 4.4).

It is apparent from Figure 4.4 that with respect to the station observed monthly T_{max} values, the HadRM3 grids consistently under-estimate mean T_{max} for most of the spring, summer and autumn months. While for the winter months there is a variable over-estimation of mean T_{max}.

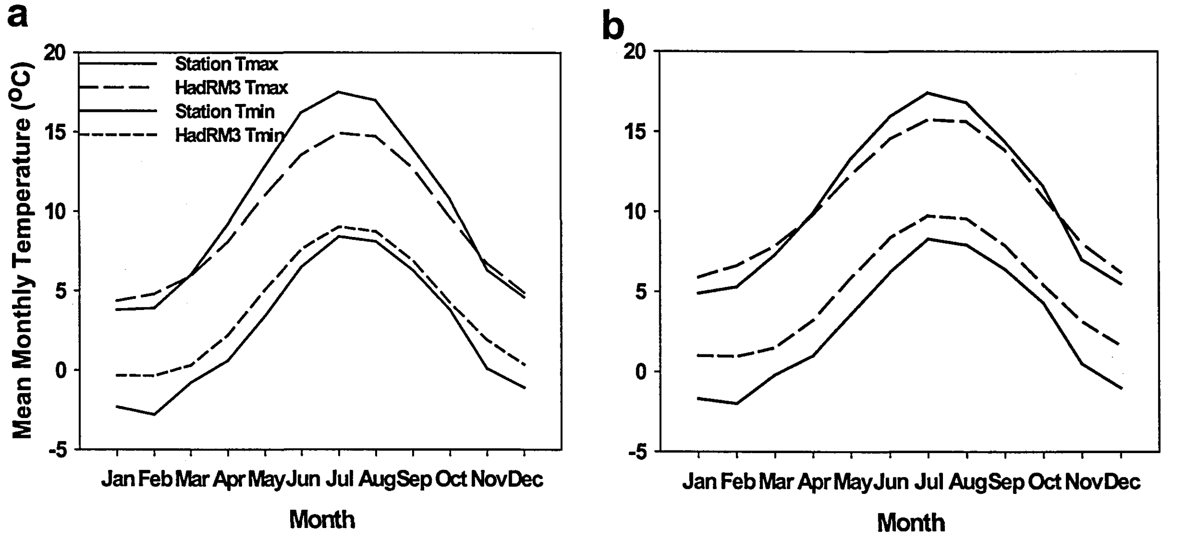


Figure 4.4: Mean monthly T_{max} and T_{min} values for the 1961-1990 baseline period. (a) Braemar station and Had Grid 145 simulated; (b) Kinbrace station and Had Grid 110 simulated.

Whereas for the estimation of T_{min}, there appears to be a more consistent warm bias in the parameterisations of the HadRM3 grids which is most marked in the late autumn to early spring months at both stations. Taken together with the T_{max} simulated values for these months, these suggest an overall warm bias for winter and early spring in the HadRM3 temperature simulations. By contrast, the greatest cold bias for T_{max} in the HadRM3 grids is associated with the summer months, particularly for Had Grid 145. These results tend to support the interpretation offered in Section 4.1.2 for the seasonally resolved data in relation to local topographic controls on temperature for these seasons.

4.2 Station Observed And HadRM3 Simulated 1961-1990 Precipitation Comparisons

4.2.1 Method I: Island sites included in analysis

Following the procedures described in Section 2.4.4.3, the HadRM3 grid-cell simulated mean seasonal 1961-1990 Precip values were subtracted from all fifty five unadjusted station values and the differentials plotted (Figure 4.5). Pearson correlation coefficient (r) and probability values (p) were computed between the HadRM3 simulated and observed 1961-1990 values for Precip (Table 4.5).

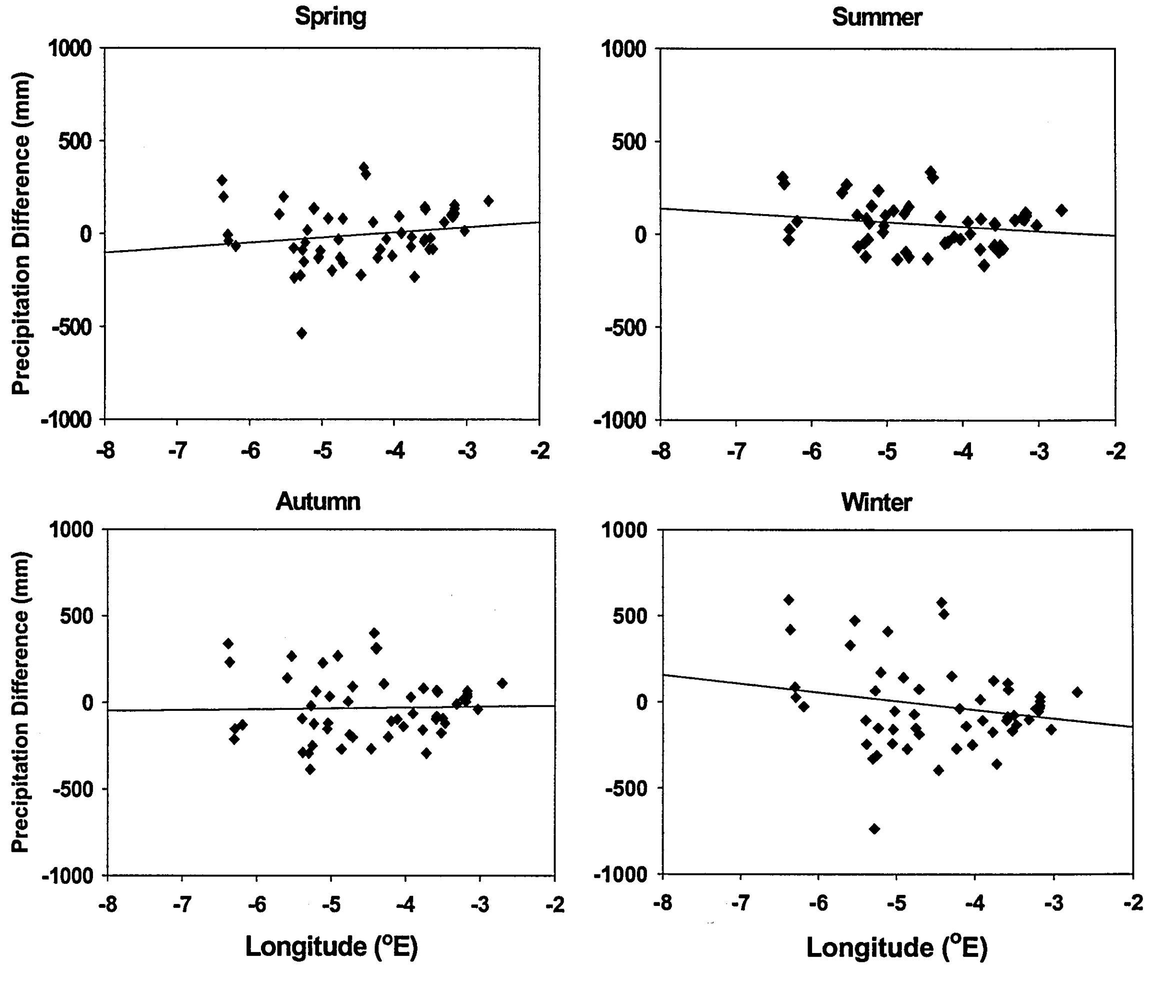


Figure 4.5: Precipitation differences (mm) HadRM3 – observed station values (n = 55).

Figure 4.5 illustrates that for all stations and by comparison with T_{max} and T_{min} for example, there is not such a clear west to east gradient in Precip differences between the stations and the corresponding HadRM3 grid cells. The least scatter in Precip totals are associated with stations in the east, whereas the biggest scatter in Precip totals is associated with western and inland stations coincident with mountainous terrain for all seasons. As was noted for temperature in section 4.1.1, at least part of this pattern is attributable to the closer concordance of station and HadRM3 grid elevations in the east. However, while some of these disparities are associated with the elevation mismatch between the HadRM3 grids and the stations, the magnitude of some of the differences also suggest parameterisation problems with some of the HadRM3 grid cells.

Table 4.5: Pearson correlation coefficient (r) and probability values (p) – comparison of station observed and HadRM3 simulated mean seasonal precipitation values

Variable	Spring	Summer	Autumn	Winter
Mean Precip	$r = 0.515$, $p < 0.001$	$r = 0.617$, $p < 0.001$	$r = 0.677$, $p =< 0.001$	$r = 0.672$, $p < 0.001$

Correlations are strongest in autumn ($r = 0.677$) and weakest in spring ($r = 0.515$), while the correlation coefficients are statistically significant ($p < 0.001$) across the seasons (Table 4.5). Given the highly localised controls governing precipitation variability in mountainous regions, it is not that surprising that the HadRM3 grid cells are producing a variable pattern of over and under representation with respect to observed values across the region. Despite progressive increases in model resolution, it is not yet possible to resolve atmospheric processes and their interaction with topography in sufficient detail to predict rainfall behaviour (Huntingford *et al.* 2003).

Based on this preliminary analysis, results would tend to suggest that this is certainly the case for the Highlands.

4.2.2 Method II: Island sites excluded from analysis

Following the procedures described in Section 2.4.4.4 including the infilling of observed values and the dropping of island sites, the exercise described in 4.2.1 above was repeated and the results recorded (Figure 4.6; Table 4.6).

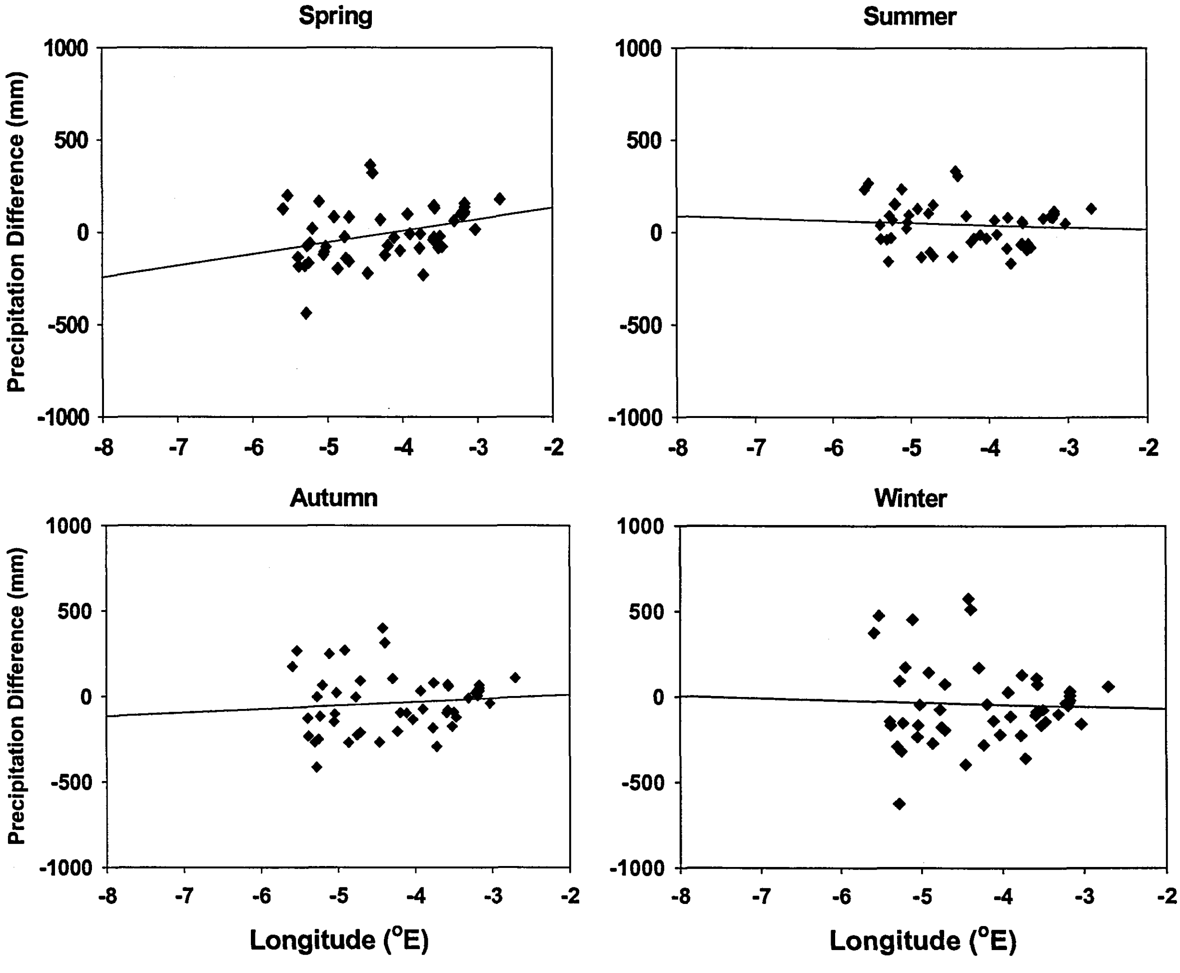


Figure 4.6: Precipitation differences (mm) HadRM3 – observed station values (n = 50), Pearson correlation coefficient (r) and probability value (p)

Following the dropping of the island sites from this analysis and the infilling of values where possible. Both the reduced correlation coefficients obtained across the seasons ($r = 0.483 – 0.643$, $p<0.001$; Table 4.6) and the pattern of scatter (Figure 4.6) suggest that the greatest differences in simulated and observed Precip is associated with central mountainous areas. This refinement to the approach would also continue to support an interpretation based on a parameterisation problem with some of the HadRM3 grids and some of the comments offered in Section 4.2.1 above.

Table 4.6: Pearson correlation coefficient (r) and probability values (p) – comparison of station observed and HadRM3 simulated mean seasonal precipitation values

Variable	Spring	Summer	Autumn	Winter
Mean Precip	$r = 0.483$, $p < 0.001$	$r = 0.538$, $p < 0.001$	$r = 0.597$, $p =< 0.001$	$r = 0.643$, $p < 0.001$

Figure 4.6 illustrates a pattern of variable over and under-representation of seasonal precipitation west to east. However, for all seasons there is more residual scatter associated with western and inland stations than for stations further east. Over-estimation at high grid box average elevations has been reported elsewhere for HadRM3, particularly for winter and spring months (Fowler *et al.*, 2005). While annual total anomalies arising from underestimates in summer and autumn precipitation have also been reported, particularly for HadRM3 Scottish grid cells (Fowler *et al.*, 2005).

Taken together, these results suggest that HadRM3 largely exaggerates the observed west-east rainfall gradient across the UK. These overestimates of mean annual rainfall in areas of high (HadRM3 model) elevation contrast with very low simulated values of mean annual rainfall in leeward areas and are thought to be the result of an over-strong orographic control within the model (Fowler *et al.*, 2005: Frei *et al.*, 2003).

Certainly for the Highlands, the substantial under-estimation of seasonal precipitation associated with HadRM3 grid ID 126 is known to be the result of an over-strong rain shadow effect within the model (pers comm UKCIP02, 2002; R. Jones, 2003). The outlying values associated with this are particularly evident in the spring and winter plots in Figure 4.5. With reference to the results recorded in Figure 4.5, it is perhaps significant that Fowler *et al.* (2005) report the largest over and under-estimations of mean annual rainfall in HadRM3 are associated with the west coast and central and eastern areas of Scotland respectively.

Despite these findings, results reported here do not lend themselves to such a straightforward interpretation. Fowler *et al.* (2005) do not explicitly report any orographic correction for the daily rain gauge data used in their Scottish regions. Similarly, they do not report on the elevation of the stations used in their North of Scotland (NS) sub-set. Given some of the complications arising from the elevation mismatch between observed station data and the HadRM3 grids noted elsewhere in this work, perhaps this indicates their findings should be interpreted with some caution. Issues surrounding corrections for orography and how these might be tackled are discussed and explored in more detail in Section 4.2.3 below.

However, these sorts of problems in the simulation of precipitation are not confined to HadRM3. In their inter-comparison study of outputs from five RCMs and observed data (daily precipitation) in an Alpine study region, Frei *et al.* (2003) found;

- a number of the models tended to overestimate the amplitude of the annual cycle, showing for example excessive dryness in late summer and early autumn (for HadRM3 the under-estimate is as large as 35% in August and September);

- excessive rain shadowing in the lee of topographic barriers associated with the overestimation of topographic precipitation enhancements on the windward side of mountain regions.

However, by way of illustrating contrasting results, when a high resolution physically-based downscaling scheme (as opposed to a nested RCM), was embedded within a GCM and run for mountain regions globally, the absence of a treatment for sub-grid rain shadows resulted in too little precipitation being simulated on the windward side of mountain ranges and too much precipitation being simulated on the leeward side (Ghan *et al.*, 2006).

These technical problems associated with RCM simulations of the baseline climate only add to the uncertainties associated with regional climate change projection. This nesting of high resolution RCMs within the GCMs introduces a further source of uncertainty associated with how robustly such models are able to downscale global projections to national and finer scales (Rowell, 2004). Given that the complex topography and land-sea distribution of the Highlands are responsible for mesoscale flow features and precipitation processes in response to seasonally variable synoptic scale disturbances, it is not surprising that HadRM3 and the driving GCM are not capturing the spatial variability of precipitation here.

4.2.3 Method III: Applying a component of orographic correction selectively for purposes of inter-comparison

Following from the procedures described in Section 2.4.4.5, station data were areally averaged under selected HadRM3 grid cells west to east (Had Grid ID's 125, 144, 163; Figure 2.1). Since it is generally accepted that grid box outputs from climate models have the spatial characteristics of areal averages (Osborn and Hulme, 1997;

Frei et al, 2003; Fowler *et al.*, 2005), on this basis it is assumed that each RCM grid represents a 50km climate. However, the evaluation of precipitation statistics is confronted with the complication that observations and model data differ in areal representativity (Frei *et al.*, 2003; Jenkins and Lowe, 2003).

Rain gauge measurements are point observations influenced by site-specific conditions and their high-frequency statistics carries the signal of small-scale variations of the precipitation systems compared to the area-mean value over an RCM grid box (Frei *et al.*, 2003). Therefore the extent to which this may be expected to truly represent precipitation over an area is not clear and may depend for example on the degree of numerical diffusion and the effective resolution of the model topography (Frei *et al.*, 2003).

Consequently, the validation of climate model simulations require observed datasets which are both historically and geographically extensive (Osborn and Hulme, 1997). However, as has been previously noted these are not available for substantial tracts of the Highlands. Consequently, the detail of even the average annual rainfall distribution from observed station data cannot be mapped with confidence (Weston and Roy, 1994; Roy, 1997). While the complex local controls on climate here and particularly local orographic controls on precipitation make interpolation of available datasets problematic.

Nonetheless, for some of the HadRM3 grid cells a number of observed station records (n = 5-9) with a mean elevation which is at least reasonably close to the corresponding HadRM3 grid can be identified. These are summarised in Table 4.7 below;

Table 4.7: Summary of HadRM3 grid cell elevations and areally averaged station elevations

HadRM3 Grid ID	Grid Elevation	Corresponding Station IDs	Mean Station Elevation (m)
125	384.65	45, 49, 51, 52, 54	291.40
144	412.19	14, 15, 27, 28, 41	436.2
163	339.39	2, 3, 4, 5, 6, 7, 8, 9, 10	315.78

Obvious from Table 4.7 is that the mismatch in mean observed station elevations (HadRM3 – stations) and the corresponding HadRM3 grids range from –24.2 m for HadRM3 grid 144 to +93.6 m for HadRM3 grid 125. With no correction for orography, the seasonal differences (HadRM3 simulated – mean station observed) are summarised in Figure 4.7a.

Despite difficulties surrounding the spatial interpolation and altitudinal projection of datasets (see Section 4.3), a relatively dense network of gauges around some of the major hydro-electric and water supply schemes together with some historical observations from Ben Nevis provide useful data on the variation of rainfall with height (Roy, 1997). Close to the west coast, the rate of increase is about 240 mm/100 m, whereas over the Cairngorms the increase is about 120 mm/100 m (Weston and Roy, 1994; Roy, 1997).

In order to keep the level for adjustment to orography to a minimum (and hence the associated assumptions), these values were only applied to the areally averaged station data and to the nearest metre beneath the corresponding HadRM3 grid. Thus the mean seasonal station values for the stations corresponding to the HadRM3 grids recorded in Table 4.3 were adjusted by;

- +93 metres, HadRM3 grid 125; -24 metres, HadRM3 Grid 144; +24 metres, HadRM3 grid 163. *Pro-rata* corrections of 240 mm/100 m were applied to the western stations beneath HadRM3 grid 125, while corrections of 120 mm/100 m were applied to the eastern station values corresponding to HadRM3 grids 144 and 163.

The values corrected for orography are summarised in Figure 4.7b below and are presented alongside the unadjusted values (Figure 4.7a) to allow easier inter-comparison;

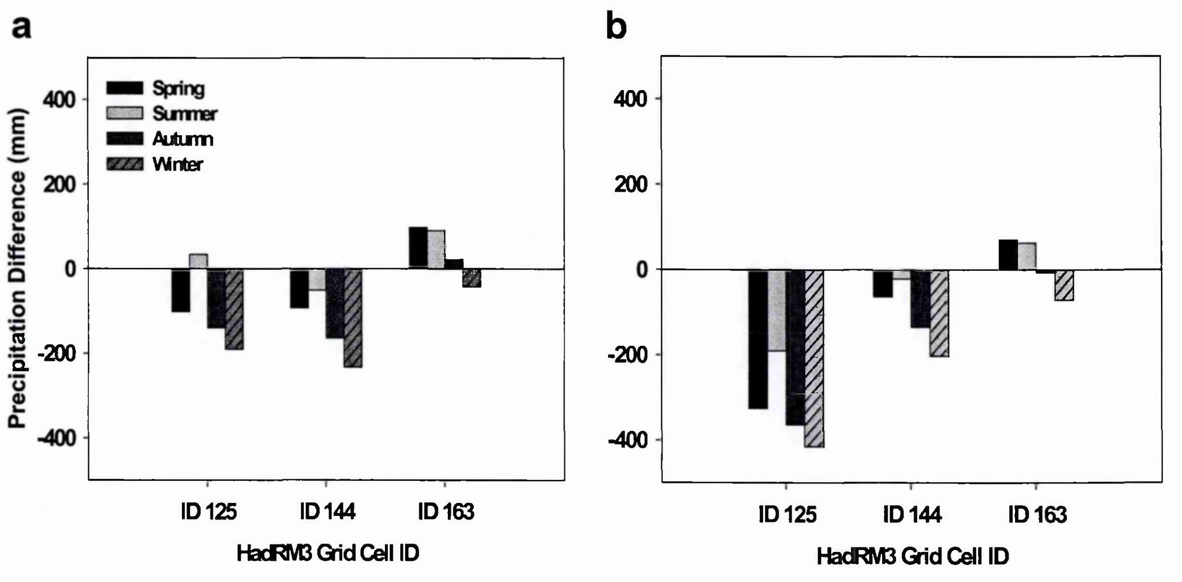


Figure 4.7: Precipitation differences (mm), selected HadRM3 grids – areally averaged station values (n = 5-9). (a) Unadjusted for orography; (b) Adjusted for orography.

Figure 4.7a illustrates that when unadjusted areally averaged station data are used, the HadRM3 grids corresponding to mountainous locations in the north west (ID 125) and inland (ID 144) largely under-represent mean seasonal precipitation with the exception of summer for HadRM3 grid 125 where there is a slight over-representation. While the largest under-representations are associated with autumn and winter mean totals for both of these grids (Figure 4.7a). By contrast, the HadRM3 grid (ID 163) associated with a station grouping furthest to the east over-represents spring, summer and autumn precipitation and slightly under-estimates winter mean totals (Figure 4.7a).

However, when the orographic corrections are applied to the areally averaged station data, the magnitude of under and over-representation changes (Figure 4.7b). When the station values for the more eastern locations corresponding to HadRM3 grids 144 and 163 are orographically adjusted, the seasonal under and over-representations associated with these locations are reduced (Figure 4.7b). Whereas this approach appears to indicate that the greatest under-representations for all seasons are associated with HadRM3 grid 125 (Figure 4.7b). Overall therefore, the exercise appears to indicate that HadRM3 is performing better for eastern locations away from the western mountains.

Taken together with the findings reported in Sections 4.2.1 and 4.2.2, it seems reasonable to conclude that the parameterisations in HadRM3, together with the lateral boundary conditions supplied by HadCM3 are not adequately capturing the spatial variation in precipitation across the region. Nonetheless, some caveats have to be applied to the numbers presented in Figure 4.7 and Table 4.4 since the values used in the orographic corrections are at best generic and do not adequately account for variation in the precipitation gradient by season.

In addition, the corrections were only applied to the mean elevation of stations beneath the corresponding HadRM3 grid cells rather than to individual point differences for each of the stations. Finally, and given discussion in preceding sections surrounding the data requirements for validation of climate model outputs, sample sizes were too small.

A further obvious point which applies to all of the above work on precipitation inter-comparison is that since baseline data for 1961 was widely missing (see previous discussion, Sections 2.2.3 and 3.2.1), a true 30-year baseline was not used. However, on a 30-year averaging period it is unlikely that one missing year would account for the magnitude of differences recorded between the observed and HadRM3 simulated baseline in the experiments detailed above.

4.3 General Discussion

While there are no consistent gradient values for precipitation enhancement with elevation, similar values have been reported elsewhere, for example;

- Harrison (1986) has reported values of 200-350mm/100 m for west coast mountains which are in accordance with values of 281mm/100 m for western areas reported by Ballantyne (1983);
- while for more sheltered eastern areas, gradients of 80-200mm/100 m (Harrison, 1986) and 88mm/100 m (Ballantyne, 1983) have been reported;
- whereas on a more Highland-wide analysis of precipitation gauges, McClatchey (1996b) reports annual gradients ranging from 97mm/100 m in the SE (Pitlochry) to 451mm/100m in the NW (Wester Ross).

Obviously within these ranges, seasonal variation in synoptic patterns allied to local topographic controls have an important influence on the gradient of altitudinal enhancement. Harrison (1986) reports a clear seasonality, with the shallowest mean gradients in late winter and early spring, whereas the steepest gradients are in autumn. Altitudinal effects on monthly precipitation totals are most marked in late autumn and winter due to the greater frequency of cyclonic storms.

Therefore these steeper gradients at this time of year coincide with the highest average rainfall and infer a strong orographic enhancement of systems already producing large rainfalls (Harrison, 1986). This should be expected since the orographic effect is most pronounced in warm fronts and in the warm sector of depressions, but is not so important in cold frontal rain events (Brunsdon *et al.*, 2001).

This also explains the shallower gradients recorded for late winter and early spring by Harrison since cold frontal rain events are likely to be more frequent in these seasons.

Similarly, in his analyses McClatchey (1996a, b) found considerable inter-annual variation in the seasonal enhancement of rainfall as well as considerable variation within sub-regions of the Highlands. This study of the rainfall-height relation across Scotland showed that the rainfall gradients are steeper in the northwest and west and lowest in the central and southeast Highlands (Mc Clatchey, 1996a). In addition there was a contrasting regional pattern in the seasonal pattern of fractional increase, with the greatest rates of increase recorded for spring and summer in the west, winter and spring in the southeast and summer in the central Highlands (McClatchey, 1996b).

In a more recent approach employing Geographically Weighted Regression (GWR), Brunsdon *et al.* (2001), encountered some problems with their statistical model in the Highlands. While in a more local study, Harrison and Clark (1998) noted that there did not appear to be a strong relationship between altitude and precipitation for the Lochaber area. However, there is a paucity of higher level stations in this area, and the orientation of the Great Glen from southwest to northeast could lead to higher low level precipitation totals than might otherwise be expected (Brunsdon *et al.*, 2001).

Despite these vexed issues surrounding obtaining locally representative values for the enhancement of precipitation with elevation, it was felt appropriate that some element of orographic correction should be used to adjust the averaged station values to the HadRM3 grid cell elevations since this would add value to the work. Consequently, the values reported for western and eastern areas by Roy (1997) above were used as a 'typical' first order approximation.

Even with high resolution RCMs providing more credible information on changes in the climate than GCMs in regions of heterogeneous terrain, uncertainties in estimates of large-scale climate change comparable to those found in GCMs are exhibited (Carter and Hulme, 1999; Jenkins and Lowe, 2003). Effectively, they will only refine the uncertainties of the climate change description produced by the GCM and not add new ones, thereby perpetuating the uncertainties originating in the large scale models (Carter and Hulme, 1999; Jenkins and Lowe, 2003). Therefore, this nesting of high resolution RCMs within the GCMs introduces a further source of uncertainty associated with how robustly such models are able to downscale global projections to national and finer scales (Rowell, 2004).

However, differences in predictions from different GCMs are generally much larger than the downscaling uncertainty. Therefore differences in results from the same RCM driven by different GCMs would be expected to be bigger than those from different RCMs driven by the same GCM, thus the largest uncertainty probably lies with the global prediction rather than the RCM downscaling (Jenkins and Lowe, 2003). As a result the uncertainties in the GCM climate change prediction are effectively cascaded, since RCMs are constrained by the same boundary conditions as the GCM used to drive them and thus the simulation results of the RCMs are strongly influenced by the selected GCM (Saelthun and Barkved, 2003).

Problems analogous to some of those noted for HadRM3, and, by extension the driving GCM, have also been recorded for the model's predecessor HadCM2. Thus for example, at the relatively coarse horizontal grid resolution of 2.5° latitude and 3.75° longitude of the HadCM2 most of the small scale precipitation patterns associated with orography and coastal processes are not evident (Jones and Reid, 2001). This point is well illustrated in Raisanen and Doscher's 1999 comparison of HadCM2 outputs with observational data for the Nordic countries. They found that the sharp but narrow observed maximum of precipitation at the western slope of the Norwegian mountains (especially evident in autumn and winter) is in the model weaker, broader, and further east (Raisanen and Doscher, 1999).

This would appear to support a widely held view that at this resolution the effects of local and regional forcings and circulation on regional climate (Section 7.5) are not captured (Christensen, 2001). Also, for the rest of the Nordic area, the simulated precipitation systematically exceeds the measured amounts, particularly in winter and spring (Raisanen and Doscher, 1999). Other findings were that HadCM2 underestimates the the average surface air temperature over the Nordic countries by 2-4°C in summer, and by a slightly smaller amount in spring; simulated total cloudiness was also generally higher (typically by 10% of full cover) than observations indicate (Raisanen and Doscher, 1999). While analysis of the atmospheric circulation around Northern Europe revealed a good overall agreement with observational data, two quantative deficiencies of note included;

- a slightly too southerly wintertime Icelandic low having too weak a notheastward extension towards the Arctic Ocean, this deficiency in the time-mean pressure field implying a weaker than observed time mean southerly flow over Scandinavia in winter;

- a general cold bias of a few °C in lower troposphere temperatures in Northern Europe in summer (Raisanen and Doscher, 1999).

A number of these systematic errors were cascaded into the first Hadley Centre RCM (HadRM2) driven by the boundary conditions of HadCM2. HadRM2 is a limited area model nested within the global HadCM2 for the greater European region, using boundary conditions such as surface pressure and horizontal wind components from the global HadCM2 model, this creating inaccuracies when the simulated and observed climates are compared (Jones and Reid, 2001). On a mesoscale level, precipitation over European land regions from HadRM2 is approximately 33% too high in comparison with observations (Jones and Reid, 2001), although orographic effects, particularly in coastal and mountainous regions are much better simulated in the finer scale RCM than in the coarser scale HadCM2 (Jones and Reid, 2001; Jenkins and Lowe, 2003).

However, in their comparison of observed and modelled precipitation, Jones and Reid (2001) found that the simulation of precipitation within the RCM is too high by comparison with observations, particularly over high topographical areas in all seasons, although less so in autumn. Additionally, the pattern of observed high threshold amounts in southwest Britain is not well simulated (Jones and Reid, 2001) suggesting there are problems with the RCM surface pressure being too high during summer, thereby not allowing the model to resolve large-scale moisture bearing frontal systems in this region (Noguer *et al.*, 1998).

Overall, results obtained here for both temperature and precipitation indicate substantive technical problems remain with climate models for a region such as the Highlands. The obvious conclusion is that if the models struggle to adequately describe the present baseline, these technical issues only undermine confidence in the projection of future climate for the region. Not least since these complications only compound some of the recognised uncertainties associated with climate change projections (Section 1.4).

5. Altitudinal Modelling of Future Changes

Synopsis

The major aims of this Chapter are:

- To refine the temperature Lapse Rate Models (LRMs) developed in Chapter 4 and validate model performance against some upland station records.
- To apply the models selectively for the altitudinal projection of selected climate change scenario outputs from HadRM3 in a representative western and eastern upland.
- To briefly explore projections of precipitation change for a selected eastern upland using selective orographic corrections

5.1 Considerations Informing Experimental Design - Altitudinal Modelling of Temperature

If changes to annual and seasonal temperature regimes are interpreted as a fundamental climatic driver, and with temperature profoundly influencing the present altitudinal zonation of communities in maritime uplands. The construction of locally resolved temperature lapse rate models (LRMs) which can be integrated with temperature change outputs from RCMs can be used to infer future shifts in the altitudinal zonings of the hills and moors. Also, given the bio-climatic differences associated with the more oceanically influenced hills of the west by comparison with the more continentally influenced eastern hills, it was considered desirable that representative site specific local models should be constructed in order that prospective differences in future changes could be evaluated and their prospective impact on vulnerable communities assessed.

While there is a shortage of observed upland station temperature records across the region, it was considered important nonetheless that the baseline LRMs were evaluated against some sort of observed record. This allowed their performance to be assessed across the range of lapse rate values applied to the annual and seasonal values of T_{max} and T_{min} (Section 5.2). The obvious reasoning was that this would engender more confidence in subsequent outputs. While it would have been preferable to have upland data records for the 1961-1990 period in order to test the LRMs against the same observational baseline, these are simply not available for many upland sites in the Highlands.

Another possible distortion arises from three of the upland data records identified (Cairnwell, Aonach Mor and Cairngorm Siesaws) having been exclusively collected in the 1990s, therefore given recent temperature trends there remains the possibility of a 'warm' bias here (see Section 3.4.1.2). However, given many of the other uncertainties and assumptions inherent in the approach and the range of temperature adjustments built into the LRM values, this remains a relatively minor concern when offset against the value of testing the models against some observed values (Section 5.2).

For purposes of evaluating the performance of the baseline LRMs, it was decided that these should be tested across the full annual and seasonal range of T_{max} and T_{min} (these results are presented graphically in Sections 5.2.1 and 5.2.2). To keep the volume of graphical outputs within reason, the lapse adjusted values for both the observed station baseline values and those for the corresponding HadRM3 grid baseline simulation are contained in the same plot.

It is considered that in the absence of any dramatic policy, social, economic or environmental changes and with climate warming occurring as indicated in the UKCIP02 scenarios, there will probably be no major noticeable effects on hill and moor habitats by the 2020s (Hamilton *et al.*, 2004). However, while there is likely to be the start of movement of constituent species and possibly some habitat fragmentation around the edges by the 2020s, there is likely to be little overall change in outward visual appearance (and altitudinal zoning) of the hills and moors (Hamilton *et al.*, 2004).

Recent bio-climatic modelling work tends to confirm this, with an indication of only a slight redistribution of montane habitats under the UKCIP02 2020s scenarios (Berry *et al.*, 2005). Nonetheless, all habitats will be on a trajectory of change which, although similar in the early stages for each emissions scenario, will result in increasingly greater ecological and visual effects as the century progresses (Hamilton *et al.*, 2004).

Consequently, and assuming that the UKCIP02 scenarios are realistic, prospective shifts in important seasonal isotherms will be both greater in altitudinal extent and more profound in terms of their prospective impact on upland communities by the 2050s and 2080s. The selective approach to scenario selection adopted here also stays within UKCIP02 guidelines, these suggest that for developing new research methodologies it may be sufficient to use just one or two scenarios to test the appropriateness of different techniques (Hulme *et al.*, 2002). Wider limitations of the approach are discussed further in Section 6.3.

Outputs presented here focus on altitudinal shifts in some key seasonal isotherms which are most likely to determine controls on plant growth, and hence possible shifts in community distribution. The specific seasonal focus is on:

- Shifts in spring minima since these will determine the altitudinal band which will become frost free on a mean basis. In addition, any shifts in the 0°C isotherm will control the extent and duration of late lying snow and through this critically impact communities dependant on snow for thermal protection against spring frost.
- Concomitant shifts in mean maxima associated with warming are likely to determine potential shifts in the altitudinal bands at which growth and flowering for certain species can occur. In terms of inferring shifts of species currently associated with lower vegetation zones into higher ones, shifts in both the T_{max} and T_{min} values presently associated with the upper limit are likely to be critical isotherms. Summer mean temperature changes are of particular interest, since summer mean temperatures are observed to correlate with treeline position over large scales (Korner, 1999; Grace *et al.*, 2002; Moen *et al.*, 2004).
- Shifts in winter mean minima in order to assess any prospective altitudinal shifts in the mean frost free zone. Shifts in mean maxima with altitude, particularly the 0°C isotherm since this will largely determine the altitudinal band where future snow cover may be expected.

5.2 Evaluating Lapse Rate Model (LRM) Performance Against Observed Upland Records

While upland temperature station records are scarce across the region, there are intact records available for some years at representative sites across the region, ranging from Aonach Mor in Lochaber, via Cairnwell in the Eastern Grampians, to the two records on Cairngorm (Table 5.1). Therefore the derived mean annual and seasonal maxima and minima (Table 5.2) for each of these stations in the altitudinal range 663 metres - 1245 metres were used to assess the performance of the LRMs.

Table 5.1: Upland station details and data record lengths

Station and Elevation	Data Record Notes
Cairngorm Chairlift, 663m	Winter = 1982-1998 (17yrs). Annual values and other seasons = 1981-1998 (18yrs).
Cairnwell, 933m	Winter = 1996-1999 (4yrs). Annual values and other seasons = 1995-1999 (5yrs).
Aonach Mor, 1033m	Winter = 1993-1999 (7yrs). Annual values and other seasons = 1992-1999 (8yrs).
Cairngorm Siesaws, 1245m	Winter = 1993-1999 (7yrs). Annual values and other seasons = 1992-1999 (8yrs).

Table 5.2: Derived mean annual and seasonal T_{max} and T_{min} by station

Station and Elevation	T-Value	Annual	Spring	Summer	Autumn	Winter
1. Cairngorm Chairlift, 663m	T_{max}	8.41	7.22	14.57	8.51	3.25
	T_{min}	2.29	0.59	7.43	2.96	-1.83
2. Cairnwell, 933m	T_{max}	3.68	1.80	8.72	4.49	-0.50
	T_{min}	0.82	-1.02	5.58	1.78	-3.28
3. Aonach Mor, 1033m	T_{max}	3.12	1.21	7.71	3.52	-1.08
	T_{min}	0.03	-1.89	4.55	0.46	-4.12
4. Cairngorm Siesaws, 1245m	T_{max}	1.92	0.72	5.81	2.44	-2.31
	T_{min}	-0.88	-2.19	3.08	-0.30	-5.41

The annual and seasonal station data for the upland stations reflect dramatic changes in the thermal environment with altitude in maritime upland areas and emphasise just how sharply differentiated in meteorological terms the highest hills are by comparison with the surrounding valleys. For example when the results for Aonach Mor (Table 5.2) are compared with nearby Onich (Appendix Tables 3.3 and 3.4), these emphasise the interactions of topography and landscape in determining climate heterogeneity. Here, even within the same geographical area, differences in site exposure between exposed hillsides and deep glens produce very different temperature regimes.

This is especially true above 900 metres where the range of both spring and winter maxima and minima in particular become notably suppressed by comparison with surrounding lowland sites (Table 5.2). For example, the mean winter maxima of -1.08°C on Aonach Mor is very different from a mean winter maxima of 6.83°C at Onich (15 metres). Similarly, a mean spring minima of 3.48°C at Onich is in marked contrast to the -1.89°C recorded on Aonach Mor. Even for Aonach Mor in the more oceanic west, there are not large departures in absolute values above 900 metres for these seasons compared to the slightly higher elevation Cairngorm site in the more continental east (Table 5.2). For instance, the mean spring minima are -1.89°C and -2.19°C respectively at each location.

5.2.1 Onich station values projected altitudinally

In this section the LRMs were used to altitudinally project 1961-1990 mean temperature baseline values for;

- Onich as a representative west coast station.
- The HadRM3 grid cell (ID 160) corresponding to Onich's location (mean elevation 460.54 m).

The performance of the models in projecting both sets of baseline values are then compared with the four upland observed station records (Figures 5.1 and 5.2).

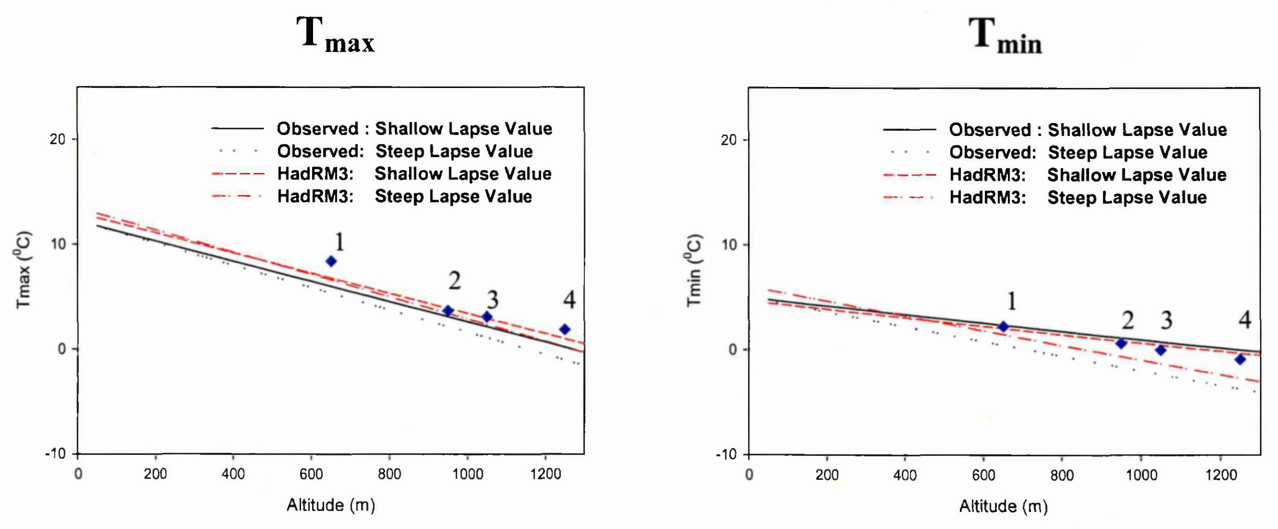


Figure 5.1: Onich station observed 1961-90 and HadRM3 simulated 1961-90 (denoted in red) mean annual T_{max} and T_{min} values adjusted altitudinally. Blue diamonds denote observed upland station values, see Table 5.2.

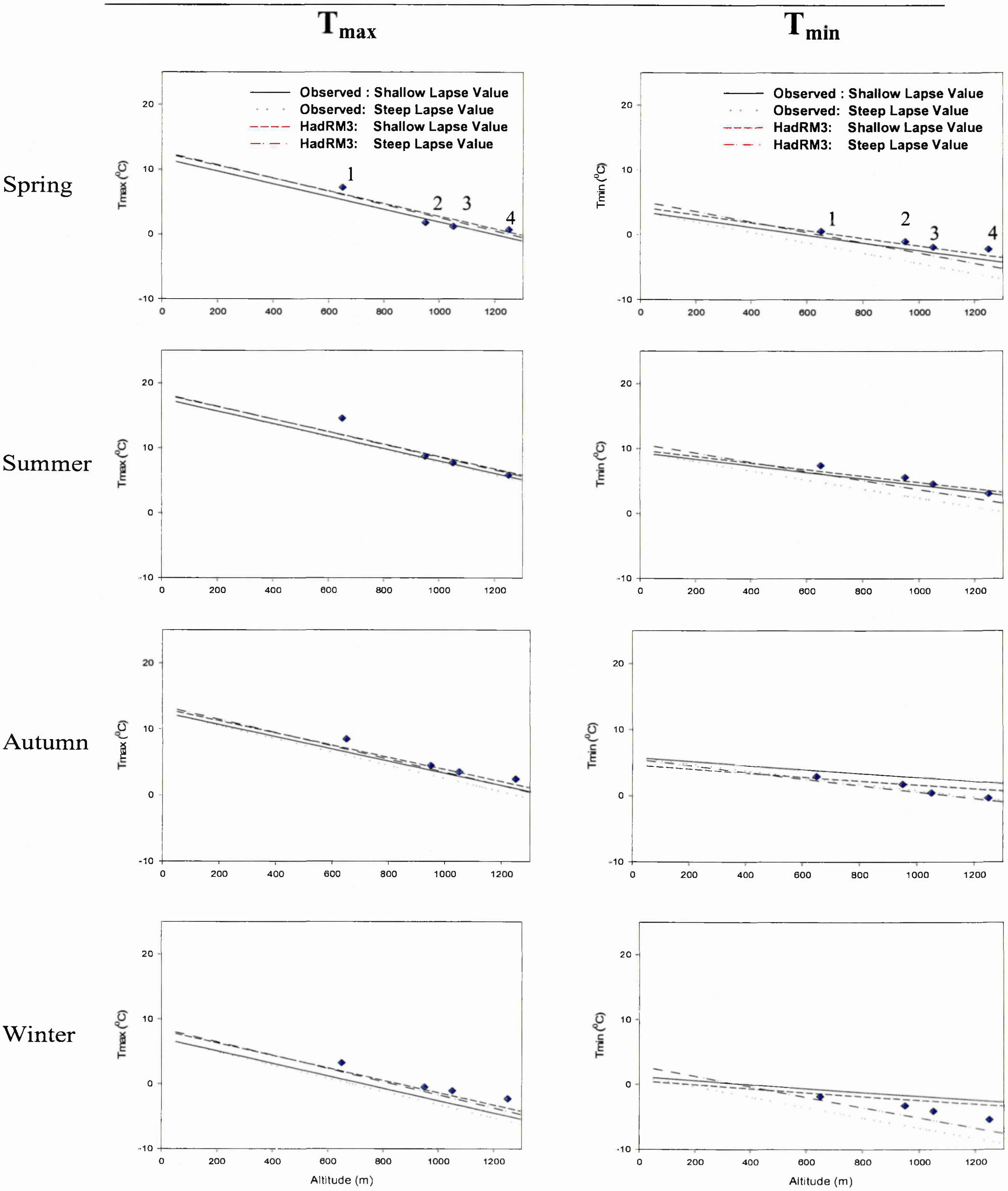


Figure 5.2: Onich station observed 1961-90 and HadRM3 simulated 1961-90 (denoted in red) mean seasonal T_{max} and T_{min} lapse adjusted altitudinally. Blue diamonds denote observed upland station values, see Table 5.2.

The full annual and seasonal range of lapse rate values (Table 2.3) were used in constructing the models in order to assess their performance against the observed station records (Figures 5.1 and 5.2). With reference to the mean seasonal station values for T_{max}, there is a cold bias in in the Onich values projected via the LRMs (Table 5.3). That is, there is an under-representation in the LRM-derived values with respect to the upland station observed values. This is least marked for the summer and spring values and most marked in autumn and winter (Figure 5.2, Table 5.3).

Similarly, for the Onich spring T_{min} projected values there is a cold bias (Table 5.3). However, the wider range of lapse rate values used for T_{min} (Table 2.3) ensure that the (broad by comparison with T_{max}) projections fall within the range of observed values for summer autumn and winter (Figure 5.2, Table 5.3).

Table 5.3: Lapse rate model range for projected Onich baseline values of T_{max} and T_{min}. Upland station values (nearest corresponding elevation) are bracketed for comparison.

Lapse Rate Model Projections	**Mean Seasonal Temperature Range (°C)**			
	Spring	**Summer**	**Autumn**	**Winter**
T_{max}				
650 m	5.28 - 5.30 (7.22)	11.21 - 11.34 (14.57)	6.02 - 6.53 (8.51)	0.35 - 0.74 (3.25)
950 m	2.34 - 2.36 (1.80)	8.27 - 8.46 (8.72)	3.02 - 3.77 (4.49)	-2.71 - -2.14 (-0.50)
1050 m	1.36 - 1.38 (1.21)	7.29 - 7.50 (7.71)	2.02 - 2.85 (3.52)	-3.73 - -3.1 (-1.08)
1250 m	-0.6- -0.58 (0.72)	5.33 - 5.58 (5.81)	0.02 - 1.01 (2.44)	-5.77- -5.02 (-2.31)
T_{min}				
650 m	-1.6- -0.33 (0.59)	4.83 - 6.1 (7.43)	2.57 - 3.84 (2.96)	-3.95- -0.78 (-1.83)
950 m	-2.13- - 1.63 (-1.02)	2.73 - 4.6 (5.58)	1.07 - 2.94 (1.78)	-6.35- -1.68 (-3.28)
1050m	-2.73- - -2.33 (-1.89)	2.03 - 4.1 (4.55)	0.57 - 2.64 (0.46)	-7.15- -1.98 (-4.12)
1250m	-3.93- -3.73 (-2.19)	0.63 - 3.10 (3.08)	-0.43 - 2.04 (-0.30)	-8.75- -2.58 (-5.41)

The systematic disparities between the projected and HadRM3 simulated values can be explained with reference to the summary information in Table 5.4. The HadRM3 grid cell (ID 160) is realised at 460.54 metres compared to the Onich station at 15 metres. Since the LRMs were built at 50 metre vertical increments, both station and HadRM3grid values were lapse adjusted to a common elevation of 450 metres to allow an inter-comparison of values. Station mean temperature values were subtracted from the HadRM3 grid mean temperature values and summarised (ΔT°C, Table 5.4).

Table 5.4: Summary of Onich observed 1961-90 values and HadRM3 grid cell 160 lapse adjusted to 450 metres for mean annual and seasonal T_{max} and T_{min}. 1 and 2 denote values derived from the application of shallow and steep lapse rate values respectively. These obtained from the range presented in Table 2.3.

Temp.	Annual		Spring		Summer		Autumn		Winter	
(°C)	1	2	1	2	1	2	1	2	1	2
Observed Max.	7.91	7.47	7.26	7.24	13.26	13.17	8.37	8.02	2.66	2.39
Had 160 Max.	8.73	8.75	8.15	8.14	13.95	13.95	8.94	8.95	3.89	3.89
ΔT°C	0.82	1.28	0.89	0.9	0.69	0.78	0.57	0.93	1.23	1.5
Observed Min.	3.18	1.87	0.87	0	7.1	6.23	4.44	3.57	-0.18	-2.35
Had 160 Min	2.89	2.92	1.58	1.6	7.52	7.54	3.3	3.32	-0.82	-0.77
ΔT°C	-0.29	1.05	0.71	1.6	0.42	1.31	-1.14	-0.25	-0.64	1.58

Across the range of annual and seasonal maxima it is clear that HadRM3 consistently over represents these values with respect to the lapse adjusted Onich values. This warm bias is particularly evident in winter, hence the consistent and systematic differences observed in Figures 5.1 and 5.2.

However, some care must be exercised in this interpretation due the sizeable elevation adjustment (435 metres) applied to the Onich station values using the LRM and the cold biases in the LRM projected values. While LRM performance can be validated against the various observed station records above 600 metres, no station data have been used at lower elevations to validate performance in the lower elevation bands.

Observational data for these intermediate elevations are not available for west coast locations as the bulk of station records are at or close to sea level. While data are available for Dall Burn (232 metres), Braemar (339 metres) and Balmoral (283 metres). These are eastern sites with a more continental climate and subject to more pronounced localised frost hollow effects in winter and spring by comparison with Onich (Section 3.1.1). It was considered therefore that they would not constitute a valid test of the model's performance on the west coast.

By contrast, across the range of annual and seasonal T_{min} there is a less consistent pattern between both sets of lapse adjusted values. For instance, with reference to the station values (ΔT°C, Table 5.4), HadRM3 appears to be simulating a spring T_{min} range which is too high (+0.71°C, +1.6°C) by comparison with the Onich observed values projected to 450 metres. Whereas for autumn T_{min}, HadRM3 appears to be simulating lower values (-1.14°C, -0.25°C) compared to the projected Onich values.

Again however, some care must be exercised in this interpretation due to the sizeable elevation adjustment applied to the observed station values and the wider range of lapse adjustment values used in the 50 metre increments for T_{min}. The range of differences for winter minima (-0.64°C, +1.58°C) underline this point since the station observed values have been lapse adjusted through 435 metres using a relatively wide range of lapse values, 0.15°C/50 metres (shallow) to 0.4°C/50 metres (steep) (Table 2.5).

5.2.2 Balmoral station values projected altitudinally

In this section the LRMs were used to altitudinally project 1961-1990 mean temperature baseline values for:

- Balmoral as a representative station in the eastern Grampians.
- The HadRM3 grid cell (ID 145) corresponding to Balmoral's location (mean elevation 320.73 m).

The performance of the models in projecting both sets of baseline values compared with the four upland observed station records with the four upland observed station records (Figures 5.3 and 5.4).

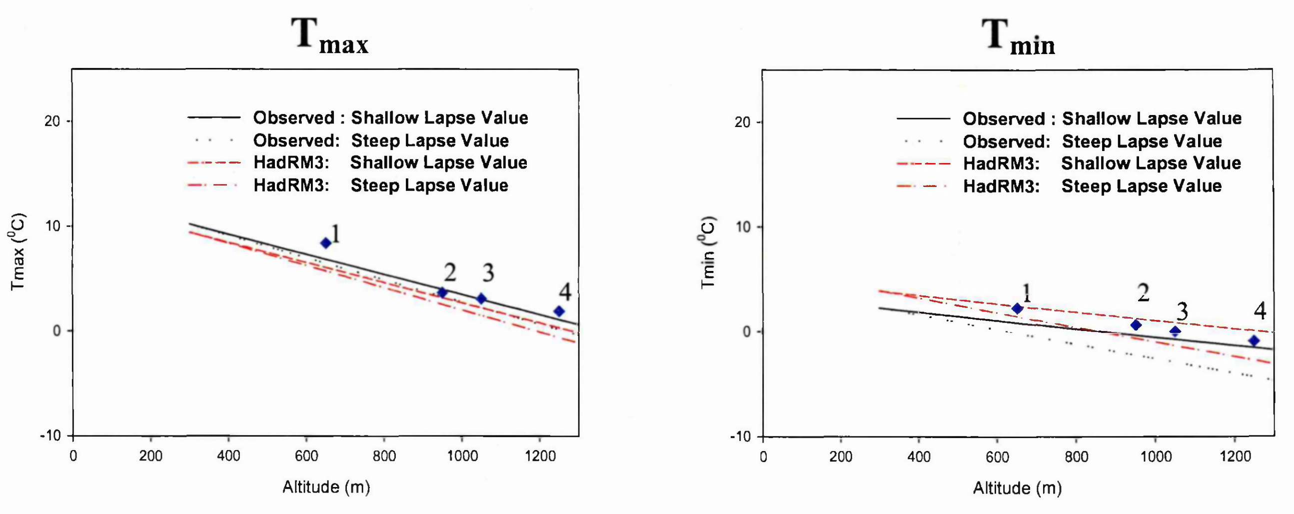


Figure 5.3: Balmoral station observed 1961-90 and HadRM3 simulated 1961-90 (denoted in red) mean annual T_{max} and T_{min} lapse adjusted altitudinally. Blue diamonds denote observed upland station values, see Table 5.2.

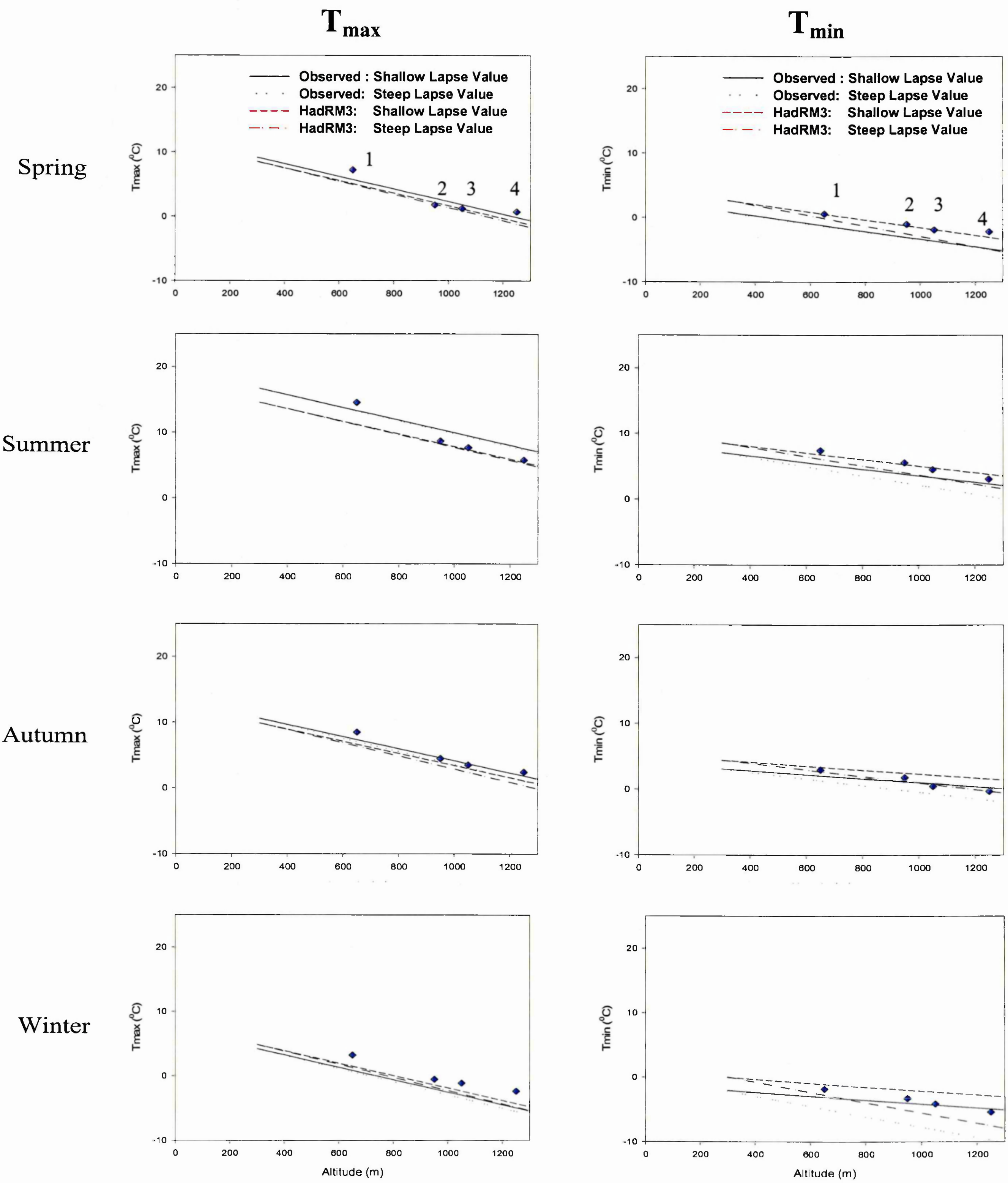


Figure 5.4: Balmoral station observed 1961-90 and HadRM3 simulated 1961-90 (denoted in red) mean seasonal T_{max} and T_{min} lapse adjusted altitudinally. Blue diamonds denote observed upland station values, see Table 5.2.

As was the case with the Onich projected values, the full annual and seasonal range of lapse rate values (Table 2.3) were used in model construction and to assess LRM performance against the observed station records by season. As with Onich, when compared to the mean seasonal station values, there is a cold bias at higher elevations in the LRM projected Balmoral T_{max} values, with the obvious exception of the summer warm bias (Table 5.5). However, by comparison with the Onich projected values (Table 5.3, Figure 5.2), the cold bias disparities associated with projected seasonal T_{max} are somewhat less (Table 5.5, Figure 5.4).

Table 5.5: Lapse rate model range for projected Balmoral baseline values of T_{max} and T_{min}. Upland station values (nearest corresponding elevation) are bracketed for comparison.

Lapse Rate Model Projections	**Mean Seasonal Temperature Range (°C)**			
	Spring	**Summer**	**Autumn**	**Winter**
T_{max}				
650 m	5.73-5.74 (7.22)	12.77-12.86 (14.57)	7.08-7.37 (8.51)	0.65-.087 (3.25)
950 m	2.79-2.80 (1.80)	10.32-10.46 (8.72)	4.08-4.61 (4.49)	-2.41- -2.01 (-0.50)
1050 m	1.81-1.82 (1.21)	9.34-9.50 (7.71)	3.08-3.69 (3.52)	-3.43- -2.97 (-1.08)
1250 m	-0.15- -0.14 (0.72)	7.38-7.58 (5.81)	1.08-1.85 (2.44)	-4.96- -4.41 (-2.31)
T_{min}				
650 m	-1.34- -1.24 (0.59)	4.63-5.35 (7.43)	1.31-2.05 (2.96)	-4.95- -3.11 (-1.83)
950 m	-3.14- -3.04 (-1.02)	2.53-3.85 (5.58)	-0.19-1.15 (1.78)	-7.35- -4.01 (-3.28)
1050m	-3.74- -3.64 (-1.89)	1.83-3.35 (4.55)	-0.69-0.85 (0.46)	-8.15- -4.31 (-4.12)
1250m	-4.94- -4.84 (-2.19)	0.43-2.35 (3.08)	-1.69-0.25 (-0.30)	-9.75- -4.91 (-5.41)

Similarly, for projected T_{min} values there is a cold bias (Table 5.5). However, the wider lapse rate range used for T_{min} (Table 2.3) deliver a projected T_{min} value for higher elevations that is within the station observed range for autumn and winter (Table 5.5). However this is not the case for the spring and summer values of projected T_{min} (Table 5.5, Figure 5.4).

Mean winter T_{max} for both the projected Balmoral observed values and the HadRM3 simulated values have a notable cold bias in relation to the station values (Table 5.5, Figure 5.4). This is likely to be a reflection of the meteorological characteristics of the site for the projected station values (subject to frost hollow effects). Whereas the HadRM3 projected values are carrying through the winter warm bias (by comparison with observed values) for HadRM3 grid cell 145 (Section 4.1.3).

Much of Deeside is a frost-hollow situated in a natural bowl in the mountains and subject to the ponding of cold air from the surrounding hills, particularly under anticyclonic conditions in winter (Section 3.1.1). This is evident in the observed winter T_{min} values (Figure 5.4). However, the wider range of values integrated into the 50 metre increments used for minima result in the lower (shallower lapse) value being closer to the station observed values when projected upslope.

The systematic disparities between the projected observed and HadRM3 simulated values can be explained with reference to the summary information in Table 5.4. For this location the HadRM3 grid cell (ID 145) is realised at 320.72 metres, whereas the Balmoral station is at 283 metres. In this case the lapse rate values were used to adjust both station and HadRM3 values to a common elevation of 300 metres for an inter-comparison of respective values (Table 5.4).

Station mean temperature values were subtracted from the HadRM3 grid mean temperature values and summarised ($\Delta T°C$, Table 5.6). Unlike the projected Onich baseline values where a substantial adjustment had to be made via the LRMs (15 metres projected to 450 metres) to adjust station values to the elevation of the HadRM3 grid. For this location, the closer accordance of the station elevation to that of the HadRM3 grid allows for a less ambiguous interpretation of HadRM3's performance in relation to the observed values (Section 4.1.3 and Figure 4.4a [Braemar monthly resolved data]). With reference to the station values (ΔT°C, Table 5.6) HadRM3 consistently under-estimates annual, spring, summer and autumn mean maxima, whereas the simulation of winter maxima are slightly higher (Section 4.1.3).

The differentials for summer maxima are particularly marked (-2.12°C and -2.1°C) and strongly suggest that the parameterisations within HadRM3 do not adequately capture localised valley heating effects in this part of the Highlands (Section 4.1.3 and Figure 4.4a). These systematic disparities explain the offset difference in the altitudinally projected HadRM3 values by comparison with the observed in Figures 5.3 and 5.4. By contrast, with reference to the station values (ΔT°C, Table 5.6) HadRM3 substantially over-represents mean T_{min} across the full annual and seasonal range (Figure 4.4a).

The differences for winter and spring T_{min} are particularly marked (+2.01°C, +2.2°C and +1.78°C, +1.86°C respectively). These disparities suggest that HadRM3's parameterisations are not adequately capturing the frost hollow effects at this particular site (Sections 4.1.2 and 4.1.3) and the effects of sometimes late lying snow into spring (Section 3.1.1). Again, these systematic disparities explain the offset differences in the altitudinally projected HadRM3 values by comparison with those for the Balmoral station in Figures 5.3 and 5.4.

Table 5.6: Summary of Balmoral observed 1961-90 values and HadRM3 grid cell 145 lapse adjusted to 300 metres for annual and seasonal T_{max} and $T_{min.}$ 1 and 2 denote values derived from the application of shallow and steep lapse rate values respectively. These obtained from the range presented in Table 2.3.

Temp.	**Annual**		**Spring**		**Summer**		**Autumn**		**Winter**	
(°C)	**1**	**2**	**1**	**2**	**1**	**2**	**1**	**2**	**1**	**2**
Observed Max.	10.22	10.2	9.17	9.17	16.7	16.69	10.59	10.58	4.23	4.22
Had 145 Max.	9.46	9.48	8.52	8.52	14.58	14.59	9.87	9.89	4.87	4.88
ΔT°C	-0.76	-0.72	-0.65	-0.65	-2.12	-2.1	-0.72	-0.69	0.64	0.66
Observed Min.	2.27	2.33	0.86	0.82	7.1	7.08	3.1	3.06	-2.06	-2.15
Had 145 Min	3.89	3.95	2.64	2.68	8.56	8.6	4.4	4.44	-0.05	0.05
ΔT°C	1.62	1.62	1.78	1.86	1.46	1.52	1.3	1.38	2.01	2.2

5.3 Utilising The LRMs To Evaluate Future Upslope Migration Of The Mean Seasonal 0°C Isotherm

In this section the LRM 1961-1990 station observed baseline values were perturbed by HadRM3 outputs of ΔT°C from the corresponding grid cell for the 2050s Medium-Low Scenario (HadRM3 grid cell ID's are 160 and 145 for Onich and Balmoral respectively). In subsequent sections (5.3.2 – 5.3.4), this approach is repeated utilising HadRM3 ΔT°C outputs for the 2050s Medium-Low Scenario and 2080s High scenario.

5.3.1 2050s Medium-Low changes to mean winter T_{max} – Onich and Balmoral

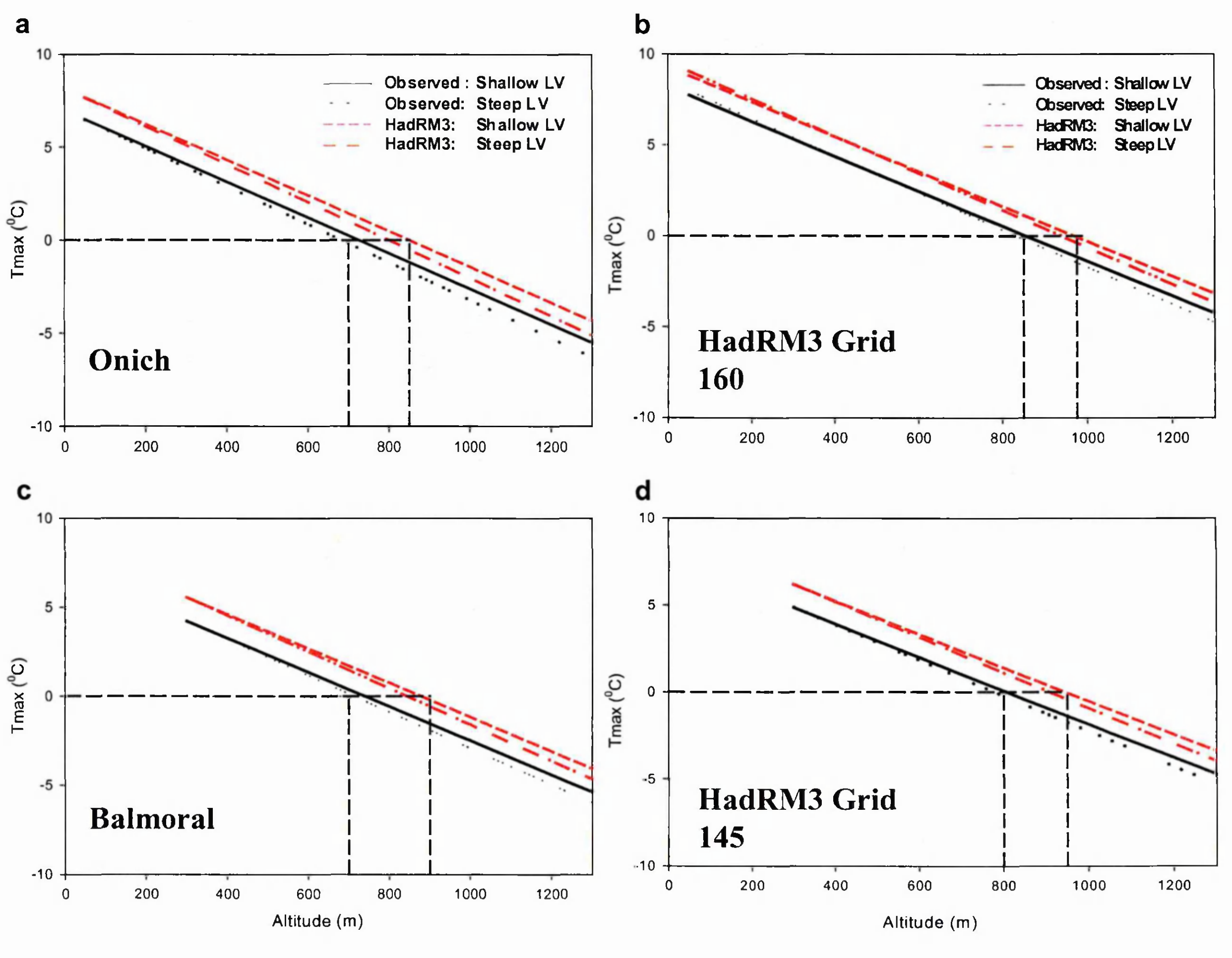


Figure 5.5: HadRM3 2050s outputs (denoted in red) applied to the Onich 1961-90 observed baseline (denoted in black) (a) and to the HadRM3 grid 160 simulated 1961-90 baseline (b). Applied to the Balmoral 1961-90 observed baseline (c) and to the HadRM3 grid 145 simulated 1961-90 baseline (d)

HadRM3 grid 160's over-estimation of mean winter T_{max} for the 1961-90 simulation by comparison with the Onich observed 1961-90 values are apparent in Figure 5.5 and continue through when the 2050s values are applied to the LRMs. As a consequence, HadRM3's simulation of the altitudinal band associated with the 0°C isotherm is higher for both the 1961-90 period (850 metres) and for when the 2050s outputs are applied (975 metres).

By comparison, the Onich observed values indicate the 0°C isotherm band is at ~700 metres for the projected 1961-90 values, this reflecting the lower station baseline means by comparison with the HadRM3 grid simulated baseline. This lower estimate of a possible future shift is propagated through the LRM when the HadRM3 2050s outputs are applied, with the 0°C isotherm shifting to 850 metres.

HadRM3 grid 145's over-estimation of mean winter T_{max} for the 1961-1990 baseline simulation are also apparent in Figure 5.5 when compared to the Balmoral 1961-1990 observed values and this systematic offset due to the HadRM3 cell warm bias is again apparent when the LRM is perturbed by the 2050s values. Consequently, HadRM3's simulation of the altitudinal band associated with the projected 0°C isotherm is higher for both the 1961-90 period (800 metres) and for when the 2050s outputs are applied (950 metres).

By comparison, the Balmoral observed values indicate the 0°C isotherm band is at ~750 metres for the projected 1961-1990 values. Again, this lower estimate is propagated through the LRM when HadRM3 2050s outputs are applied, with a shift to 900 metres.

5.3.2 2080s High changes to mean winter T_{max} – Onich and Balmoral

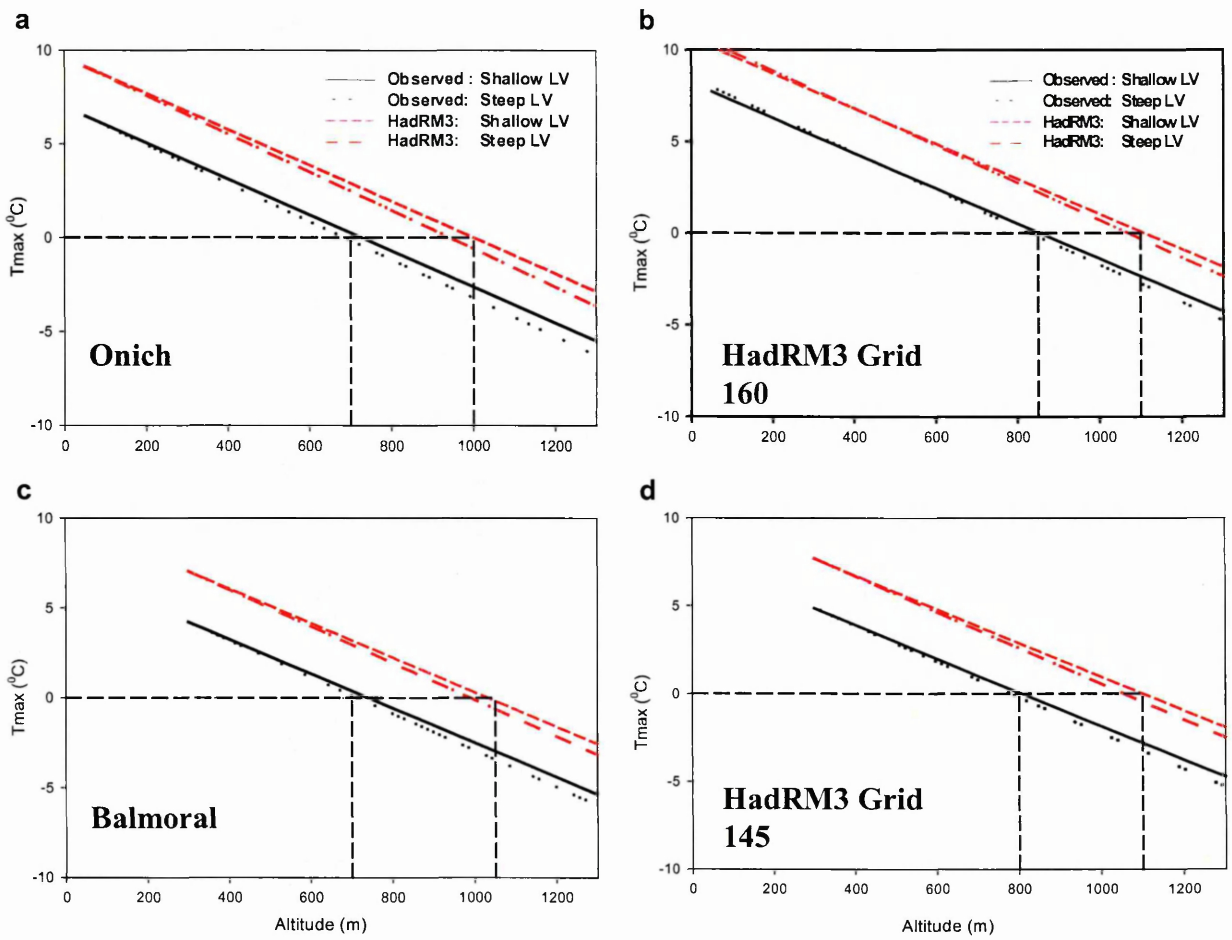


Figure 5.6: HadRM3 2080s outputs (denoted in red) applied to the Onich 1961-90 observed baseline (denoted in black) (a) and to the HadRM3 grid 160 simulated 1961-90 baseline (b). Applied to the Balmoral 1961-90 observed baseline (c) and to the HadRM3 grid 145 simulated 1961-90 baseline (d). LV denotes lapse value.

HadRM3 grid 160's over-estimation of mean winter T_{max} for the 1961-1990 simulation by comparison with the Onich observed 1961-1990 values are apparent in Figure 5.6 and the warm bias is projected through when the 2080s values are applied to the LRMs. As a result, HadRM3's simulation of the altitudinal band associated with the 0°C isotherm is higher for both the 1961-90 period (850 metres) and for when the 2080s outputs are applied (1100 metres). Whereas the Onich observed values indicate the 0°C isotherm band is at 750 metres for the projected 1961-1990 values. This lower estimate is propagated through the LRM when the HadRM3 2080s outputs are applied, with the 0°C isotherm shifting to 1000 metres.

The over-estimation of mean T_{max} associated with the HadRM3 grid 145 1961-1990 baseline simulation are also apparent in Figure 5.6 by comparison with the Balmoral values and the warm bias is evident when the LRMs are perturbed by the 2080s values. Consequently, HadRM3's simulation of the altitudinal band associated with the projected 0°C isotherm is higher for both the 1961-90 period (800 metres) and for when the 2080s outputs are applied (1100 metres). Whereas the Balmoral observed values indicate the 0°C isotherm band is at 700 metres for the projected 1961-1990 values. Again, the lower values associated with the projected station baseline is propagated through the LRMs when HadRM3 2080s outputs are applied, with a shift to 1050 metres.

5.3.3 2050s Medium-Low changes to mean spring T_{min} – Onich and Balmoral

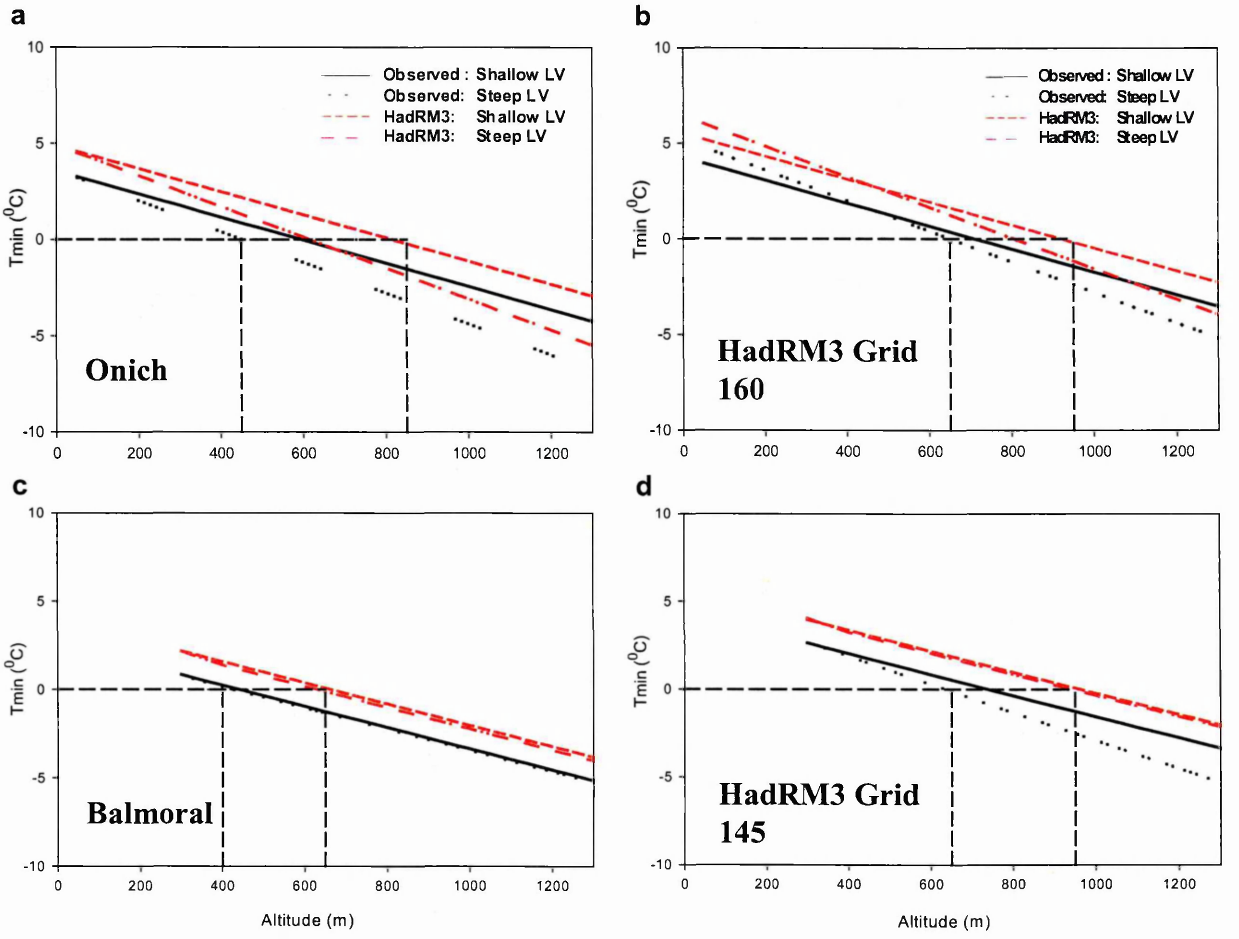


Figure 5.7: HadRM3 2050s outputs (denoted in red) applied to the Onich 1961-90 observed baseline (denoted in black) (a) and to the HadRM3 grid 160 simulated 1961-90 baseline (b). Applied to the Balmoral 1961-90 observed baseline (c) and to the HadRM3 grid 145 simulated 1961-90 baseline (d)

The over estimation of mean spring T_{min} for the 1961-1990 simulation by HadRM3 grid 160 are apparent in Figure 5.7 when compared to Onich. With the warm biases again carried through when the LRMs are perturbed by 2050s outputs. Consequently, HadRM3's simulation of the altitudinal band associated with the 0°C isotherm is higher for both the 1961-90 period (650 metres) and for when the 2050s outputs are applied (950 metres). Whereas the Onich observed values indicate the 0°C isotherm band is at 450 metres for the projected 1961-1990 values and this lower estimate (associated with lower station values) is propagated through the LRM when the HadRM3 2050s outputs are applied, with a shift in 0°C isotherm to 850 metres.

HadRM3 grid 145's over-estimation of mean T_{min} for the 1961-1990 baseline simulation are also apparent in Figure 5.7 compared to the Balmoral and this warm bias is again propagated through when the LRM is perturbed by the 2050s values. Consequently, HadRM3's simulation of the altitudinal band associated with the projected 0°C isotherm is higher for both the 1961-1990 period (650 metres) and for when the 2050s outputs are applied (950 metres).

By comparison, the Balmoral observed values indicate the 0°C isotherm band is at 450 metres for the projected 1961-90 values. Again, the lower estimate associated with the projected colder station values is apparent when HadRM3 2050s outputs are applied, with a shift to 650 metres.

5.3.4 2080s High changes to mean spring T_{min} – Onich and Balmoral

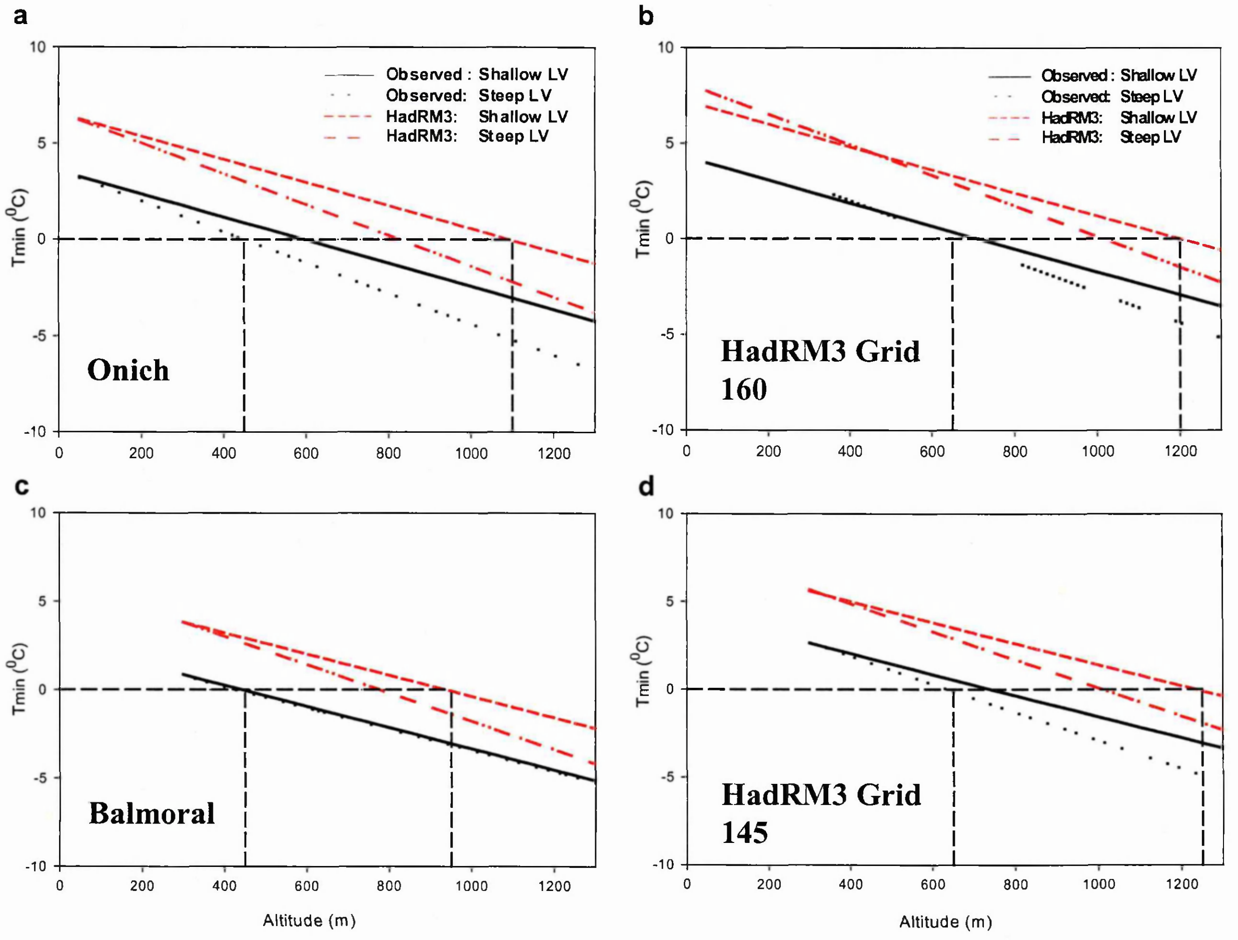


Figure 5.8: HadRM3 2080s outputs (denoted in red) applied to the Onich 1961-1990 observed baseline (denoted in black) (a) and to the HadRM3 grid 160 simulated 1961-1990 baseline (b). Applied to the Balmoral 1961-1990 observed baseline (c) and to the HadRM3 grid 145 simulated 1961-1990 baseline (d)

The warm bias for spring minima associated with HadRM3 grid 160 are apparent when compared with the Onich values in Figure 5.8. Consequently, HadRM3's simulation of the altitudinal band associated with the 0°C isotherm is higher for both the 1961-1990 period (650 metres) and for when the 2080s outputs are applied (1250 metres). By comparison, the Onich observed values indicate the 0°C isotherm band is at 450 metres for the projected 1961-1990 values. This lower estimate associated with lower station values is propagated through the LRM when the HadRM3 2080s outputs are applied, with a shift in the 0°C isotherm to 1100 metres.

Figure 5.8 again illustrates HadRM3 grid 145's over-estimation of mean minima for the 1961-1990 baseline simulation compared to Balmoral. The warm bias is again apparent when the LRM is perturbed by the 2080s values. As a consequence, HadRM3's simulation of the altitudinal band associated with the projected 0°C isotherm is higher for both the 1961-90 period (650 metres) and for when the 2080s outputs are applied (1250 metres). By comparison, the Balmoral observed values indicate the 0°C isotherm band is at 450 metres for the projected 1961-1990 values. This lower estimate is projected through the LRM when HadRM3 2080s outputs are applied, with a shift to 950 metres.

The LRM-derived projections of shifts in the 0°C isotherm are explored further in a case study context in Section 6.2, while a wider use of the LRMs to infer thermally driven shifts in upland vegetation zones is more fully explored in Section 6.1.

5.4 Orographic Projections Of Precipitation Change For A Selected Eastern Upland

This section extends the approach explored in Section 4.2.5, while also taking account of some of the issues discussed in Section 4.3 relating to the application of corrections for orography. Therefore for purposes of consistency, the same Weston and Roy (1994) enhancements 'typical' of an eastern upland (120 mm/100 m) are applied to the areally averaged station data (n = 9) corresponding to HadRM3 grid 163 described in Table 4.7. The approach was applied to selected seasons since the greatest future summer drying and winter wetting relative to 1961-90 are simulated by HadRM3 grids corresponding to the eastern Highlands, such as grid 163. In addition, and as was previously explored in Section 4.2.5 the simulation of the 1961-90 baseline by HadRM3 grid ID 163 most closely matches the values obtained when the observed station values are inter-compared with the HadRM3 grid simulations (Figure 4.7).

A further advantage of targeting this location is that a considerable number of designated communities are located in these eastern uplands e.g. Consequently, HadRM3 simulated summer drying allied to temperature changes means drought stress will have considerable implications for habitats of high conservation value e.g. Similarly, changes to winter precipitation patterns allied to temperature changes have significant implications for communities associated with late lying snow e.g. Therefore, and as described in Section 2.6, the Weston and Roy adjustments are used in combination with perturbation by $\Delta\%$ Precip changes from HadRM3 grid 163 to project future changes to 500 m and 1000 m respectively for the selected scenarios (Figure 5.9).

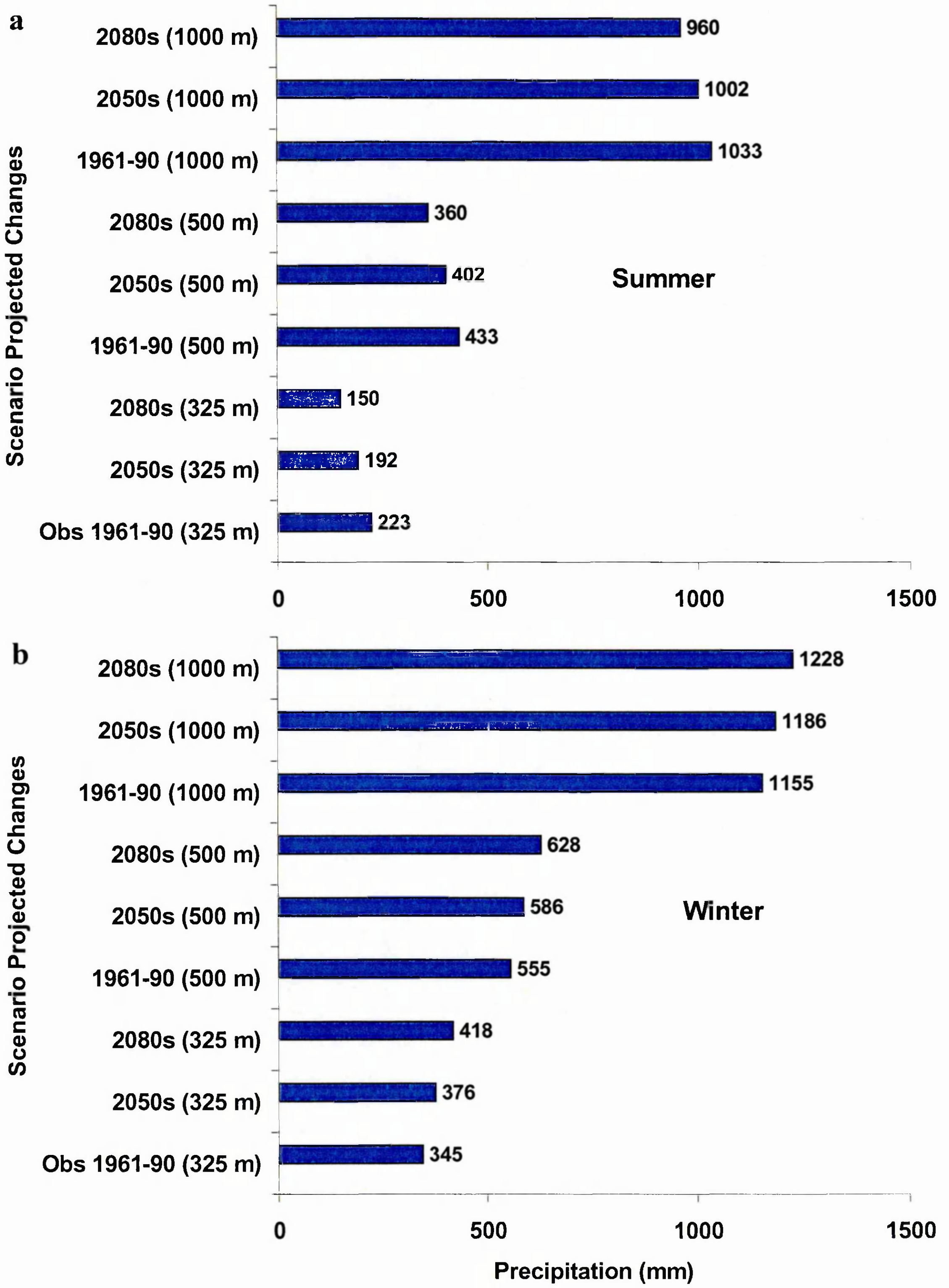


Figure 5.9: Altitudinal projection of areally averaged station precipitation data; (a) summer and (b) winter for selected UKCIP02 scenarios. Observed baseline values are included for comparison.

Figure 5.9 illustrates that if the precipitation-elevation relationship is assumed to be linear, rather than curvilinear above certain elevations (Harrison, 1986). The increasing summer drying and winter wetting as the century progresses simulated by HadRM3 grid 163 can be expected to extend to upland areas.

However, some caution must be attached to this simple linear extrapolation. A study of three station records for Coire Dubh on the Isle of Rhum ranging from 21-320 metres for the 1961-90 baseline period indicated that the relationship may become curvilinear above ~300 metres and that the amplitude of precipitation fluctuation for the higher station was almost twice that of the sea-level gauge (Coll and Gil, unpublished work). However, in this study the reduced total at the 320 m gauge by comparison with the gauge at 152 m was concluded to be an artefact of measuring error due to the influence of wind at the upper site.

This wind field deformation and deflection of hydrometers over the gauge orifice is widely documented for mountain regions and results in a systematic measurement bias (Frei *et al.*, 2003), and this underestimation of precipitation in topographically complex regions plagues most gauge-based gridded precipitation datasets (Adam *et al.*, 2006). This clearly creates a further complication in using upland observational series in the evaluation of climate model performance. There are of course also considerable simplifying assumptions in the exercise presented here about future rainfall-elevation relationships in a seasonally warmer world. However, the use of climate models themselves assume that fundamental relationships within the climate system will be unchanged under altered climatic conditions, because the parameterisations they use are based on the present climate. Some of these discussion strands are explored further in subsequent chapters.

6. Applying LRM Outputs To Upland Climate Change Impact Assessments

Synopsis

The major aims of this Chapter are:

- To utilise wider LRM outputs in order to infer shifts in vegetation zones in the western and eastern mountains.
- The use of LRM modelled outputs for mean 0°C isotherm shifts to infer possible upland climate change impacts in a case study context.
- To consider the implications and limitations of applying the methods developed.

6.1 LRM derived seasonal shifts in the thermal regimes associated with present vegetation zones

For the projections undertaken in this section, approximate elevations associated with upland vegetation zones in the west and east were obtained (Figure 2.4). Poore and McVean (1957) considered there was a sub-montane zone equivalent to the Norwegian sub-alpine and forest zones and a montane zone above the potential tree-line equivalent to the Norwegian lower and middle alpine zones. While the treeline in inland southern Norway lies at between 1000 m and 1200 m (Hofgaard, 1997; Hallanaro and Usher, 2005), in an analogous fashion to the altitudinal descent of life zones in the NW Highlands associated with steep lapse rates (Angus, 2001). There is a similar descent of life zones north-westwards in Scandinavia (Thompson *et al.*, 2005), such that along the coast of northern Norway the treeline is ~200 m (Hallanaro and Usher, 2005).

While upland areas of Fennoscandia have been influenced by various human activities over a long period of time (Hofgaard, 1997; Hallanaro and Usher, 2005), zonations in these areas have been less affected than in Iceland e.g. (Thomson, 2005). Upland areas in the UK, however, have endured substantial anthropogenic impacts marked in turn by deforestation, burning and grazing management (Ratcliffe, 1977, 1990; Birks, 1988; Ratcliffe and Thompson, 1988; Thompson and Brown, 1992), consequently landscapes here appear to lack the typical sequence of altitudinal life zones found in continental Europe (Thompson and Brown, 1992; Brown *et al.*, 1993). Thus while no part of Europe's mountains are truly natural (Thompson *et al.*, 2005), anthropogenic influences on the natural ecotones are not so marked in many of the more continental high alpine areas (Pauli *et al.*, 2003) as they are in the Highlands.

Therefore, while zonations equivalent to those in Norway have been used here to better differentiate shifts in key seasonal isotherms, in terms of inferring impacts of prospective changes to thermal regimes on plant communities, these can be used interchangeably with other proposed classifications. For example, Thompson and Brown (1992) propose two upland zones based on the composition of vegetation;

- a sub-montane zone with vegetation derived mainly from woodland above the limits of enclosed farmland. For the Highlands this will vary both east to west and south to north in line with the gradation of climate;
- a montane zone equivalent to the Norwegian low-alpine zone (but with some significant absences and only a small extent of middle alpine vegetation) beginning at the potential tree line. Thus, the sub-montane zone could be classed as being under ~300 metres in the west and ~600 metres in the east, with the montane zone occurring above these elevations.

For this exercise, the shallow and steep lapse rate values (Table 2.3) were averaged for both T_{max} and T_{min} across the full altitudinal range in order to arrive at a single mean value (T_{mean}) for both T_{max} and T_{min} for each season at 50 metre increments. The mean isotherm for both T_{max} and T_{min} associated with the upper limit of each identified zone was noted for both the Onich and Balmoral 1961-1990 observed datasets. Similarly, the LRM outputs obtained when the observed 1961-1990 Onich and Balmoral baseline data were primed with HadRM3 grid outputs for the 2050s Medium-Low and 2080s High scenarios were averaged for both T_{max} and T_{min} across the 50 metre increments of the models. The mean isotherm values associated with the upper limit of each vegetation zone were then compared to the HadRM3 perturbed future values and the new elevation associated with the isotherm value noted.

The shifts in the mean isotherm for both T_{max} and T_{min} associated with the upper limits of each zone were then expressed as an upslope migration in metres for each of the scenario outputs used. Since the LRMs were built on a 50 metre increment and many of the elevation bands corresponding to the future mean isotherm values fell between these, for purposes of consistency the upper 50 metre band was used as the cut-off point for the possible future location of the isotherms.

Future isotherm shifts for each of the scenarios at Onich and Balmoral are summarised in Figures 6.1 and 6.2. However, in order to allow better scope for discussion to develop, prospective spring and autumn changes in the west and summer and winter changes in the east are examined. Obviously it is the suite of seasonal changes which will drive prospective changes in the uplands, while HadRM3 future time slice outputs project substantial increases in mean autumn T_{min} in the west and mean summer T_{max} in the east relative to 1961-1990.

6.1.1 Spring and autumn changes in the western mountains

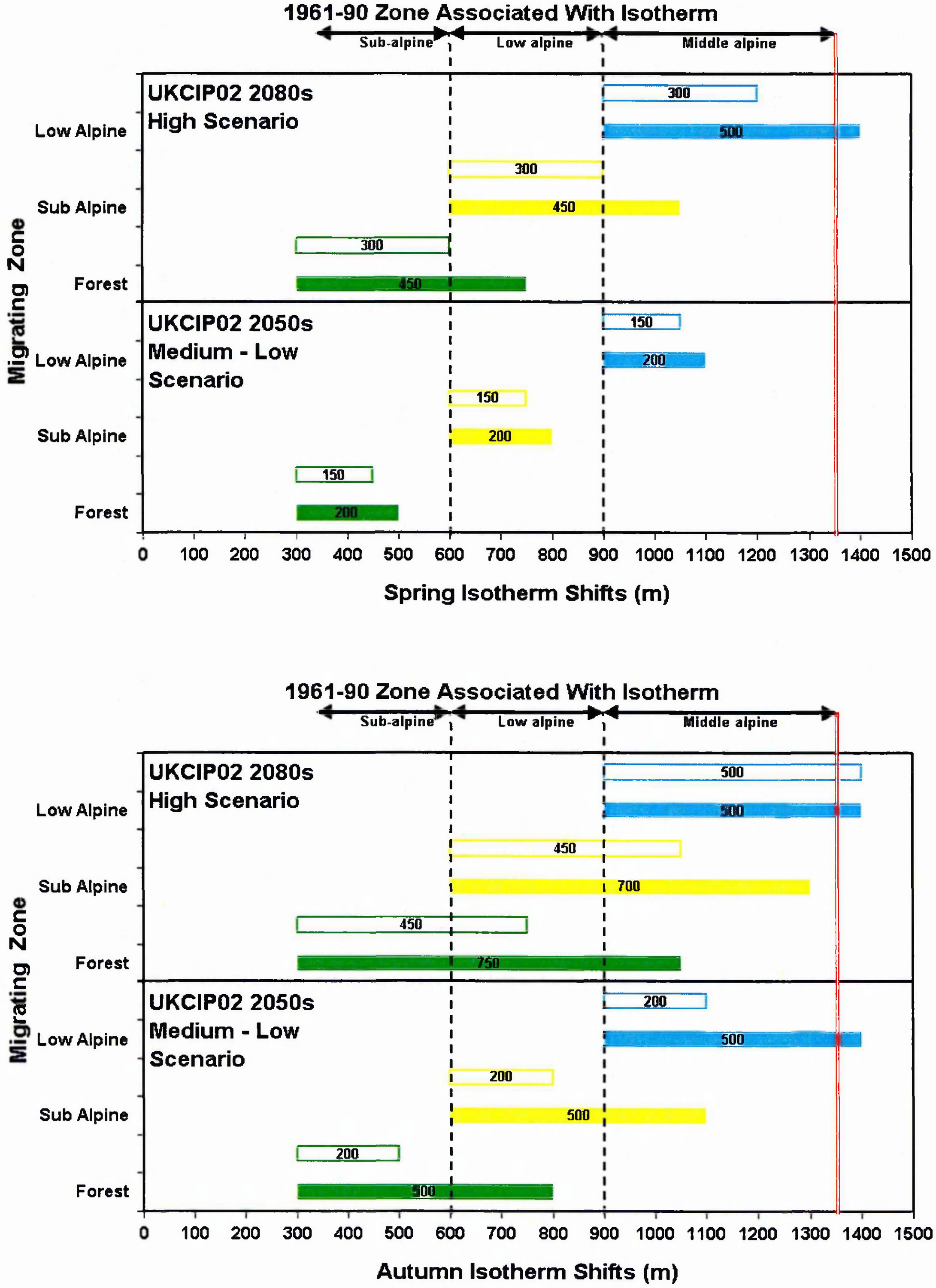


Figure 6.1: Summary of spring (upper panel) and autumn (lower panel) isotherm shifts associated with vegetation zones under selected UKCIP02 scenarios. Lower and upper blocks for each zone denote mean T_{min} and T_{max} respectively. The red line denotes the summit level.

It is apparent from Figure 6.1 that for both future scenarios, the biggest uphill shifts are associated with the mean T_{min} isotherm. The HadRM3 perturbed LRMs indicate an altitudinal migration of ~200 metres for each of the zones under the 2050s Medium-Low scenario. Whereas for the 2080s High scenario, the uphill migration of the mean T_{min} isotherm associated with each of the zones is ~450-500 metres.

However, it should be noted that the 500 metre extension of the isotherm associated with the low alpine zone into the middle alpine zone is a constrained figure. Since the LRMs were only constructed to an elevation of 1300 metres, once an isotherm associated with any zone and scenario (for both T_{min} and T_{max}) extended beyond the summits, the difference in elevation associated with the future scenario was taken to a standard cut-off point of ~1400 metres in the models. The reasoning here was pragmatic, since once the isotherm associated with any vegetation zone was above the summits it would be reasonable to infer, at least in terms of the approach here considering only thermal controls, that the associated vegetation zones potentially could be also. Therefore, while the HadRM3 derived isotherm associated with a vegetation zone could project in some cases to ~1600 - ~1800 metres, given the time involved in building the models and the time it would take to extend their altitudinal range, the exercise was irrelevant.

This approach has been consistently applied to subsequent figures. By comparison with shifts in mean T_{min}, those for mean T_{max} associated with specific vegetation zones are smaller, with the perturbed lapse rate models indicating a ~150 metre shift for each zone under the 2050s scenario, with these extending to a ~300 metre shift for each of the zones under the 2080s scenario.

The vertical migration of autumn isotherms associated with each of the vegetation zones are even more spectacular (Figure 6.1). The HadRM3 perturbed LRMs indicate a shift in mean T_{min} associated with each zone of ~500 metres under the 2050s scenario. While under the 2080s scenario the mean minimum isotherm associated with the two lower zones is extending ~700-750 metres uphill. By contrast, there is a lesser shift for mean T_{max}, with the models indicating a shift of ~200 metres for the isotherm associated with each zone under the 2050s scenario extending to ~450-500 metres (recalling the earlier caveat on the 1400 metre cut-off) under the 2080s scenario.

However, given the wider problems of incorporating uncertainty and the wide range of climate change projections from different models (Section 1.4). UKCIP02 provide a rough guide to the sensitivity margin to be applied to outputs from the HadRM3 50km x 50km grids. These are intended as a first order approximation aimed at incorporating scientific uncertainty, with a strong recommendation that the margins be applied to all simulations of UK climate change output at the 50km grid scale (Hulme *et al.*, 2002).

For the scenarios applied in this work and for average temperatures, the suggested ranges are;

- +/- 1°C of ΔT°C for average winter and summer temperatures for the Medium-Low Emissions scenario;
- +/- 2°C of ΔT°C for average winter and summer temperatures for the High Emissions scenario (Hulme *et al.*, 2002).

Since no more guidance is supplied on how these average corrections may be applied to T_{max} and T_{min} or across other seasons, these first order uncertainty corrections were applied to mean T_{max} and T_{min} across the seasons for both scenarios. The lapse rate values used in the models (Table 2.3) approximate to ~ -1.0°C/100m for mean T_{max} and ~ -0.5°C/100m for mean T_{min}.

Therefore, in terms of applying the UKCIP02 guidelines to the isotherm shifts, and since the LRMs assume linearity, these are interpreted to translate to per metre equivalents for each zone as;

- +/- 100 metres for mean T_{max} and +/- 200 metres for mean T_{min} associated with the 2050s Medium-Low scenario;
- +/- 200 metres for mean T_{max} and +/- 400 metres for mean T_{min} associated with the 2080s High scenario.

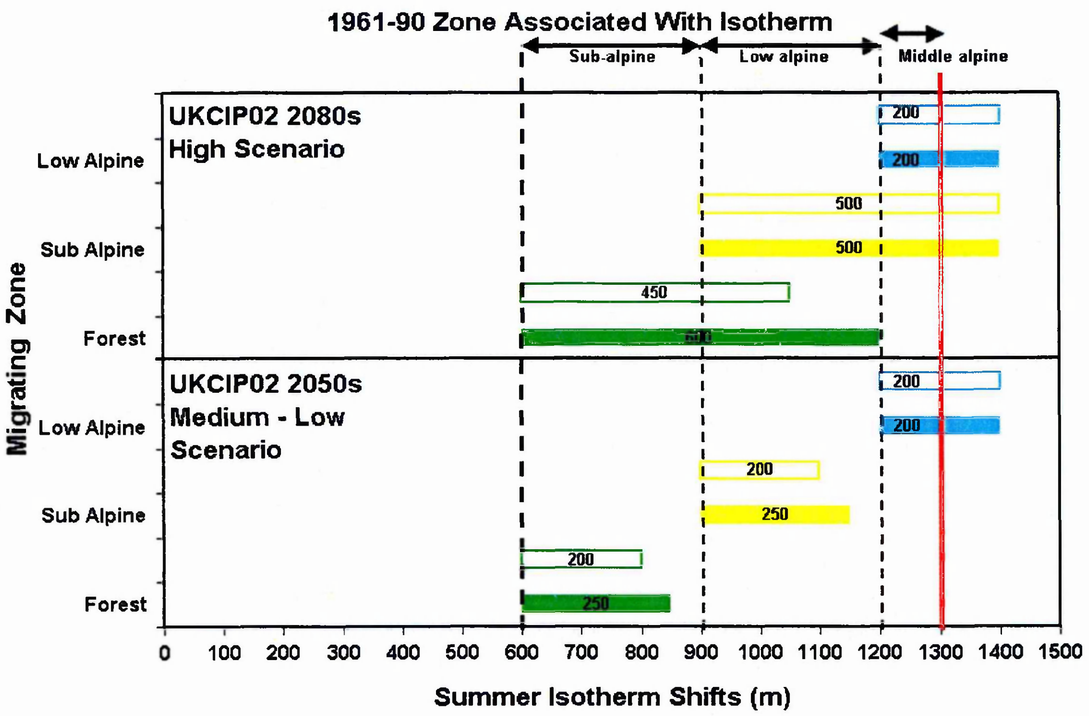
The introduction of this uncertainty approximation therefore allows a range of possible spring and autumn isotherm migrations associated with vegetation zones based on the mean shifts (Figure 6.1) to be calculated (Table 6.1).

Table 6.1: Summary of possible spring and autumn isotherm range shifts (metres) by vegetation zone in western mountains when UKCIP02 uncertainty adjustments applied. Numbers (°C) in brackets denote the 1961-90 mean isotherm values associated with the upper limit of the vegetation zone

Scenario and season	**Vegetation Zone Isotherm Range Shift (m)**		
	Forest	Sub-Alpine	Low-Alpine
Spring 2050s T_{max}	200-400m (8.72°C)	50-250m (5.78°C)	50-250m (2.84°C)
Spring 2080s T_{max}	100-500m	100-500m	Not Applied
Spring 2050s T_{min}	0-400m (1.48°C)	0-400m (-0.61°C)	0-400m (-2.71°C)
Spring 2080s T_{min}	50-800m	50-800m	Not Applied
Autumn 2050s T_{max}	100-300m (9.63°C)	100-300m (6.75°C)	100-300m (3.87°C)
Autumn 2080s T_{max}	250-650m	250-650m	Not Applied
Autumn 2050s T_{min}	300-700m (4.60°C)	300-700m (3.40°C)	300-700m (2.20°C)
Autumn 2080s T_{min}	350-1150m	300-1100m	Not Applied

It is apparent from Table 6.1 that applying the suggested adjustments create a greater range of possible shifts in T_{max} and T_{min} by the 2080s by comparison with the 2050s, this reflecting the greater uncertainty associated with warming projections for later in the century. The greater range of possible shifts in T_{min} also reflect the shallower lapse adjustment range applied in comparison to T_{max} (Table 2.3). The corrections were not applied to prospective shifts of the isotherm associated with the low alpine zone through the middle alpine zone due to the expressed shifts (in metres) associated with the 2080s being constrained figures as noted above.

6.1.2 Summer and winter changes in the eastern mountains



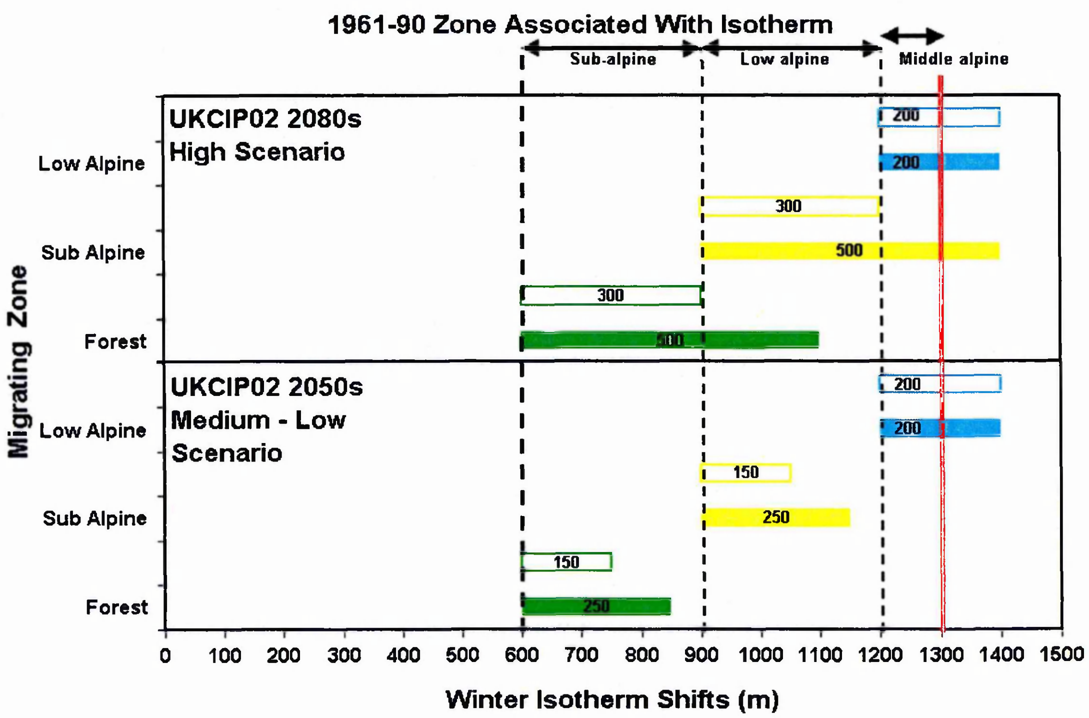


Figure 6.2: Summary of summer (upper panel) and winter (lower panel) isotherm shifts associated with vegetation zones under selected UKCIP02 scenarios. Lower and upper blocks for each zone denote mean T_{min} and T_{max} respectively. The red line denotes the summit level.

Apparent from Figure 6.2 is that for both future scenarios, the biggest uphill shifts are again for the isotherm associated with mean T_{min}. The HadRM3 perturbed LRMs indicate that these will be ~250 metres for the forest and sub-alpine zones under the 2050s Medium-Low scenario, with a ~200 metre uphill shift in the isotherm associated with the low alpine zone, again however, this is a constrained figure in light of the reasoning discussed earlier. Whereas for the 2080s High scenario, the uphill migration of the mean minimum isotherm associated with the forest zone is ~600 metres and ~500 metres for the sub-alpine zone, a shift which takes it above the summits.

Due to the cut-off line in the models being set at ~1400 metres, the shift of ~200 metres in the isotherm associated with the low alpine zone is a constrained figure. By comparison with shifts in mean T_{min}, those for mean T_{max} are smaller, with the perturbed LRMs indicating a ~200 metre shift for each zone under the 2050s scenario. These extend to a ~450 metre shift in the isotherm associated with the forest zone and a ~500 metre shift in the isotherm associated with the sub-alpine zone under the 2080s scenario. Again the ~200 metre shift in the isotherm associated with the low alpine zone is a constrained figure.

It is also apparent from Figure 6.2 that for both future scenarios, the biggest uphill shifts are again for the isotherm associated with mean minimum temperatures. The HadRM3 perturbed LRMs indicate that these will be ~250 metres for the forest and sub-alpine zones under the 2050s Medium-Low scenario, with a ~200 metre uphill shift in the isotherm associated with the low alpine zone. Again however, this is a constrained figure in light of the reasoning discussed earlier. Whereas for the 2080s

High scenario, the uphill migration of the mean minimum isotherm associated with the forest and sub-alpine zone is ~500 metres, a shift which takes the isotherm associated with the sub-alpine zone above the summits.

As before the shift of ~200 metres in the isotherm associated with the low alpine zone is a constrained figure. By comparison with shifts in mean T_{min}, those for mean T_{max} are smaller, with the perturbed lapse rate models indicating a ~150 metre shift for the isotherm associated with the forest and sub-alpine zone under the 2050s scenario. Under the 2080s scenario these extend to a ~300 metre shift in the isotherm associated with the forest and sub-alpine zone. The ~200 metre shift in the isotherm associated with the low alpine zone for both scenarios is again a constrained figure.

If the exercise aimed at incorporating a first-order approximation of scientific uncertainty (Section 6.2.1 above) is repeated. It becomes similarly possible to calculate a range of shifts based on the mean shifts (Figure 6.2) for summer and winter isotherms in the eastern mountains (Table 6.2).

Table 6.2: Summary of possible summer and winter isotherm range shifts (metres) by vegetation zone in eastern mountains for selected when UKCIP02 uncertainty adjustments applied. Numbers (°C) in brackets denote the 1961-90 mean isotherm values associated with the upper limit of the vegetation zone

Scenario and season	**Vegetation Zone Isotherm Range Shift (m)**		
	Forest	Sub-Alpine	Low-Alpine
Summer 2050s T_{max}	100-300m (13.78°C)	100-300m (10.87°C)	100-300m (7.96°C)
Summer 2080s T_{max}	250-650m	300-700m	Not Applied
Summer 2050s T_{min}	50-450m (5.29°C)	50-450m (3.49°C)	0-400m (1.69°C)
Summer 2080s T_{min}	200-1000m	100-900m	Not Applied
Winter 2050s T_{max}	50-250m (1.23°C)	50-250m (-1.71°C)	100-300m (-4.68°C)
Winter 2080s T_{max}	100-500m	100-500m	Not Applied
Winter 2050s T_{min}	50-450m (-3.75°C)	50-450m (-5.40°C)	0-400m (-7.00°C)
Winter 2080s T_{min}	100-900m	100-900m	Not Applied

It is apparent from Table 6.2 that applying the suggested adjustments creates a greater range of possible shifts in both T_{max} and T_{min} by the 2080s. These are for the same reasons discussed in Section 6.1.1 in relation to the values reported in Table 5.5 and 5.6. Similarly, the approach was not applied to shifts associated with the low alpine zone isotherm by the 2080s due to the cut-off point (metres associated with summit levels) applied to the LRMs.

Aside from the wider uncertainties associated with the RCM projections of future change, some caution must be ascribed to the outputs described here. Although the work is more locally relevant and the approach to lapse rate projection of future temperature change is considerably more refined and rigorous than, for example, that described in Moen *et al.* (2004). As indicated in Section 5.2 there are cold biases when station baseline values are projected via the LRMs for most seasons.

This implies that shifts in the mean isotherm values associated with each zone could in fact be greater than those reported here. Offsetting this, however, are concerns with how the HadRM3 parameterisations capture the 1961-1990 baseline climate. As has been demonstrated here and previously explored in Sections 4.1.2 and 4.1.3, there is a systematic seasonal T_{min} warm bias in HadRM3 grid cells. While for $T_{max,}$ the warm bias associated with winter in HadRM3 grids is countered by a cold bias in spring, summer and autumn, certainly at least for grid cells in the eastern Highlands (Figure 4.4).

Therefore, this introduces an element of further doubt surrounding the projections of future warming. These relate to the technical performance of HadRM3, since there is an inference that these biases will be incorporated in the future projections of change from the 2020s to the 2080s. This introduces an obvious concern in relation to future projections of change via the LRMs since these were primed with outputs from the RCM.

Despite the technical problems with HadRM3's performance for the region (this in addition to wider uncertainties surrounding projections of future change), and while these can be compounded by over-extrapolation of data via the LRMs. Nonetheless, and despite the difficulties, the approach delivers a more locally representative set of possible future changes to temperature with altitude than has hitherto been available for the region.

6.2 LRM derived shifts in the 0°C seasonal isotherm and case study applications

In Chapter 5 the focus was on prospective altitudinal changes in the mean occurrence of the 0°C mean winter maximum isotherm, together with prospective changes in altitude of the occurrence of the 0°C mean spring minimum isotherm. The underpinning reasoning was that these can be used as proxies for the altitude where future snow-lie may occur in the western and eastern mountains associated with warming scenarios. While all temperature and humidity changes imply associated changes in sensible and latent heat flux which affect biogeochemical cycles, and hence ecosystem processes, the 0°C threshold is particularly important in mountain areas on time scales ranging from the daily (e.g. frost cycles) to the centennial (e.g. glaciers) (Diaz *et al.*, 2003).

The location of the 0°C isotherm across the winter and spring is likely to be the most significant control on where future snow cover may be expected in the Scottish hills in these seasons. In addition any shifts in the altitudinal occurrence of this isotherm across these seasons will influence the likely duration and extent of snowlie through the late spring and early summer season. In this section the implications of possible future changes in the mean altitudinal occurrence of this isotherm are explored for;

- the ski industry as a vulnerable sector
- snowbed vegetation communities as a vulnerable habitat of high conservation value and Sub-Arctic willow scrub as one of the UK's rarest habitats with a distribution largely confined to the Highlands

A wider use of the LRMs to infer thermally driven shifts in broader upland vegetation zonations was undertaken in Section 6.2 and some of the implications and limitations of the approach are discussed further in Section 6.3, while the potential implications for the ski industry are explored in Section 6.2.1 below.

Most research into changing snow cover has emphasised the importance of weather patterns at a regional level in determining intra-annual variation, with these in turn linked to wider climatic changes at a global or hemispheric level (Harrison *et al.*, 2001). Against a backdrop of a general retreat in mountain glaciers during the 20th century, there has been a marked decrease in the spatial extent of snow cover across the northern hemisphere (Appendix Figure 6.2), particularly in spring from the 1970s onwards (IPCC, 2000; Harrison *et al.*, 2001; Morison *et al.*, 2001).

Observed changes in winter precipitation and the duration of snow cover in Scotland can be related to the anomalously high positive phase the winter NAOI has exhibited through much of the 1980s and 1990s and the associated tracking of milder, moister air across much of Northwest Europe (see Annex Paper 3). For the Scottish mountains, associated with the predominance of these Atlantic westerly systems have been winters and springs with a more ephemeral snow cover, particularly at lower elevations and occasional heavy falls in strong winds (Harrison *et al.*, 2001). The changing pattern and duration of snowlie across the 1980s reported by Harrison (1993) is also coincident with the anomalously high winter NAOI over this period (Section 3.4.2.1). While Barnett *et al.* (2006) also report a shortening of the snow season since 1961, with the greatest reductions in northern and western Scotland. Not surprisingly, this is coincident with a corresponding reduction in the number of days of frost (ground and air) and a lengthening of the growing season (Barnett *et al.*, 2006).

With most climate models simulating an increase in the winter NAOI as a regional response to greenhouse gas forcing (Gillett *et al.*, 2003; Osborn, 2005) a future increase in mean winter temperatures can be expected, and these are at least part of what is captured in HadRM3's simulation of increases in both mean winter minima and maxima across Scotland by the 2050s and 2080s. Thus, for example, in the UKCIP02 Medium-High Emissions scenario, the future trend is for an increase in the NAOI, although the year-to-year variability in the index is large. The increase in this decadal NAOI becomes significant (i.e. beyond 'natural' variability) by the 2050s (Hulme *et al.*, 2002). While in a more recent set of experiments ahead of the IPCC 4th Assessment Report, most climate models continue to indicate increasing NAOI values with increasing GHG concentrations, although with varying magnitude (Osborn, 2005).

Given the present-day relationships between the NAO and UK winter weather, this strongly suggests that future UK winters will become more westerly in nature. While the NAO linkages for some of the Highland station data demonstrated in Sections 3.4.1 and 3.4.2 and some of the possible future changes in the occupancy characteristics of NAO positive winters (Annex Paper 3) have potentially significant implications for future snow lie.

6.2.1 Potential implications for the ski industry as a vulnerable sector

Changes in the depth, duration and altitudinal extent of winter snow cover have major implications for the future viability of the Scottish ski industry. Although the sector's contribution to the UK economy is small, it is of considerable importance to the local Highland economy at a time of year when activity in other traditionally important employment sectors (e.g. farming and tourism) is slack.

For instance, during the period 1993-95 figures indicate;

- around 0.1 million trips each year were taken in Scotland purely for the purpose of skiing;
- these generating around £16 million per annum in expenditure of which around one third per annum is spent on-slope, with the remainder expended on local goods and services, as well as supporting over 1000 jobs including part-time and seasonal employment (Palutikof, 1999).

This is an industry which has already been impacted by the unreliable snow cover due to changes in the regional forcings associated with the largely positive winter NAOI observed since the 1980s.

Associated with the changes to snowfall patterns in recent decades there have been fewer operational days for ski lifts and tows and a general shortening of the ski season (Harrison *et al.*, 2001) and there is an obvious relationship between the number of days with snow lying and the number of ski days (Appendix Figure 6.3; Palutikof, 1999; Viner and Agnew, 1999). In addition, the ski areas have had to invest more time in snow grooming and fencing activities to make best use of a dwindling resource.

Together with wider economic changes and the rise of the budget airlines, overseas holidays to higher resorts with more reliable snow cover have become more affordable. Consequently the lower level ski resorts in the Highlands have become increasingly economically marginal and the winter of 2003/2004 has already seen the owners of Glencoe and Glenshee put the resorts on the market after deciding they could no longer afford to keep them open. Already therefore, for ski resorts operating with small economic margins, even relatively small climate changes (in tandem with other market forces) have had disproportionate impacts.

If the LRMs are used to calculate winter and spring (the main seasons associated with snow cover) T_{mean} ($[T_{max}+T_{min}]/2$) associated with the 2050s and 2080s scenarios respectively, the implications of the projected temperature changes at 500 m and 1000 m have significant implications for this sector. For instance if the Balmoral baseline data are used, temperature changes at both altitude bands are substantial;

- for winter T_{mean} there are increases of 1.33°C and 2.80°C at 500 m and increases of 1.29°C and 2.77°C at 1000 m compared to 1961-1990 associated with the 2050s and 2080s scenarios respectively;
- for spring T_{mean} there is an increase of 1.34°C and 2.76°C at 1000m for the 2050s and 2080s respectively compared to the 1961-1990 values.

If a 1°C increase in temperature is taken to equate to an average reduction of ~9 days snowlie (Harrison *et al.*, 2001). The above temperature increases obtained from the LRMs suggest approximate reductions in snow lie of:

- ~10+ days at 500m and 1000m under the 2050s Medium-Low scenario;
- ~20+ days at both elevation bands under the 2080s High scenario.

To place these estimates of change in context, while data on snow cover in Scotland are relatively sparse (Harrison *et al.*, 2001). Over the Scottish mountains snow typically lies for more than 50 days a year (UKMO, 2006), an estimate borne out in the analysis of Harrison *et al.* (2001), where the average number of snowlie days November-April was 61 over a 1961-1990 baseline period at Braemar. While on the summits of the Cairngorms, the average number of days with sleet or snow falling extends to over 100 (Ellis and McGowan, 2006). Appendix Figure 5.3 provides a summary of the annual number of days with snow lying at Braemar (339 m) since 1927, as well as an indication of the high inter-annual variability.

However, Harrison *et al.* (2001) also emphasise that increases in temperature operate differentially with altitude, with the gradient of change in snow cover becoming progressively less steep in a warming climate. Since the LRMs assume linearity and exhibit a cold bias at higher elevations (Section 5.2), some cautious interpretation of the above is required. However, in some recent micro-climatic modelling work in the Swiss Alps (elevation range 2050 m – 2500 m), Keller *et al.* (2005) report a comparable average decrease of a month in snowmelt for rock and sward habitats in response to a +4°C increase in mean temperature.

For an already beleaguered sector in the Scottish context, the implications are significant. Findings here support those of Harrison *et al.* (2001), namely that reliability of snow cover will be increasingly confined to slopes higher than ~1000 m as the century progresses. However, the imposition of an underlying anthropogenic trend on the natural year-to-year and decade-to-decade variability of Highland climate (Coll *et al.*, 2005), implies that some good snow winters are still likely under warming scenarios, but with diminishing frequency (Harrison *et al.*, 2001).

Consequently, available adaptation strategies for the Scottish ski industry remain similar to those proposed in earlier work (Harrison *et al.*, 2001), with a diversification into alternative activities which are less dependant on snow conditions. Unlike more continentally influenced climates elsewhere, snowmaking as a technical adaptation (Scott *et al.*, 2003; Scott and McBoyle, 2006) remains unlikely. While the possible latitudinal migration of resorts, as is a possible option elsewhere in Northern Europe (Moen and Fredman, 2006) is not a viable option here.

6.2.2 Potential implications for vulnerable plant communities of high conservation value

Changes in the altitudinal extent and duration of mean snow cover also have considerable implications for many of the arctic-alpine plant communities, particularly those reliant on snow cover for thermal protection against frost. The drifting of both falling and re-suspended snow is a major control on the spatial pattern of accumulation (Harrison *et al.*, 2001), with deep drifts most commonly formed in sheltered gullies on leeward slopes. As a result, across the Highlands the most persistent summer snowbeds are to be found on slopes with an easterly aspect, this reflecting the dominance of westerly air-streams.

Amid the concern expressed about the impact of warming on mountain vegetation, there are concerns surrounding the disappearance of snow patches and their associated vegetation in the sub-nival zone (Crawford, 1997; Klanderud and Birks, 2003). These areas of prolonged snowlie are characterised by a distinctive bryophyte vegetation (McVean and Ratcliffe, 1962). Snow patches occur over about 850 metres in many areas throughout the Highlands, with the plant communities associated with them scarce in terms of their national land cover and hence of high conservation value (Woolgrove and Woodin, 1994, 1996).

Snowbed communities in Scotland are at the extreme edge of their geographic range and are thus particularly sensitive to climate warming, particularly the winter warming predicted for north-east Britain (Woolgrove and Woodin, 1994). Watson *et al.*, (2004) report that the year 2003 had the smallest number of snow patches in north-east Scotland at the start of July since their standardised records began in 1974, with a disappearance at exceptionally early dates being a feature of a warmer and sunnier than average summer. These observations are coincident with 2003 in Scotland being the warmest in the Meteorological Office time series, as well as being the 3rd. driest and sunniest in the same series (Met. Office, 2004). While 2003 globally was the 2nd warmest in the instrumental record (CRU, 2005).

With the frequency and duration of summer snow patches strongly related to the combination of winter and spring temperatures, together with patterns of snow drift over these seasons. It is likely that the combination of increases in mean winter maxima and spring minima associated with wider warming will have profound implications for the snowbeds and their associated bryophyte communities.

With the Balmoral baseline LRMs projecting an altitudinal migration of +150 m (2050s) and +400 m (2080s) for the 0°C mean T_{max} value in winter (Section 5.3). In tandem with a vertical migration of +200 m (2050s) and +400 m (2080s) for the 0°C mean T_{min} value for spring, these would suggest dramatic changes in future community distribution. With similarly large migrations of the 0°C mean values associated with the LRM projected Onich baseline values (Section 5.3), when eastern and western changes are considered together, the implication is that there will be substantial loss of these communities across the Highlands should the level of warming associated with these climate change scenarios be realised.

The range of possible shifts in the isotherm obtained via the LRMs envelopes possible shifts reported previously by Harrison *et al.* (2001). Using the earlier UKCIP98 Medium-High scenario for the 2080s in their snowlie modelling work, they reported a possible upward shift of 250 m associated with snow related environmental zoning along mountain slopes (Harrison *et al.*, 2001).

By comparison, the greater shifts associated with the 2080s here reflect some of the differences between the UKCIP02 modelling and the earlier modelling work undertaken for UKCIP98. The UKCIP02 scenarios show a higher atmospheric concentration of CO_2 (and hence greater warming) for the High Emissions scenarios than was the case for the UKCIP98 scenarios (Hulme *et al.*, 2002).

To reinforce the above points, if Parry's (2000) estimate of a mean 1°C increase in temperature equates to a 150 m increase in snowline. If spring T_{mean} is derived from the 2050s and 2080s perturbed LRMs, changes to the 850 m T_{mean} relative to 1961-1990 are;

- +1.43°C and +1.29°C associated with the 2050s scenario for Onich and Balmoral respectively;
- +2.95°C and +2.92°C for Onich and Balmoral respectively under the 2080s scenario.

Applying Parry's figures, these suggest an upward shift of ~225 m (2050s) and ~450 m (2080s) for any snowlie presently associated with the 850 m altitudinal band. If the UKCIP02 perturbed LRM outputs are applied and linearity of temperature changes with altitude are assumed (see Section 6.1.1 above). Assuming a worst case warming scenario by the 2080s, snowpatches and their associated communities are likely to be vestigial and only associated with the highest summit areas in the Cairngorms and Lochaber.

Another community with a high conservation value and vulnerable to upland warming is Sub-Arctic willow scrub. The chamaephytic shrub *Salix lapponum* is the most widespread of the montane scrub forming species of willow found in Scotland, the principal others being *Salix aurita, Salix lanata, Salix myrsinites* and *Salix arbuscula*. McVean & Ratcliffe (1962) regarded *S. lapponum* as being the typical dominant montane scrub forming species, ranging in altitude from 220m - 1070m (Mardon, 1990). However, its natural range and habitat preferences in the Highlands are masked by the pressure of grazing, confining communities largely to cliffs, ledges and rocky slopes (Poore & McVean, 1957; Rodwell, 1992; Stewart *et al.*, 1994).

A component of the National Vegetation Classification (NVC) W20 *Salix lapponum-Luzula sylvatica* community classification, the community is largely confined to the Highlands (Rodwell, 1992). Late snow lie is especially marked in the East-Central Highlands where the *Salix-Luzula* scrub has its centre of distribution, with a marked tendency for stands to occupy sites with a North to East aspect which would afford some protection against early snow-melt (Rodwell, 1992). Emperical measures of site shelter (Harrison and Kelly, 1996) applied to sites where *S. lapponum* clumps occur in various locations in the east-central Highlands, strongly suggest an association with late-lying snow at sites largely above 500 m and with a north or north easterly aspect (Coll, 1996). *S. lanata* communities are also known to be coincident with sites associated with late lying snow patches at similar elevations and aspects (D.K. Mardon, pers. comm.). This dependence on late snow lie for thermal protection renders these communities particularly vulnerable to the combination of rising winter and spring temperatures.

Other species not found in the hyper-oceanic climate of the west and dependant on snow cover for survival such as *Luzula sylvatica, Galium saxatile* and *Blechnum spicant* are also vulnerable to changed patterns of winter and spring snowlie. The changes in the east are likely to see species dependant on snow cover, particularly beneath the highest summits displaced by those that can withstand winter exposure.

For vegetation located at an elevation where a small rise in temperature takes dormant plants out of the frozen condition even for a few days, there can be major physiological consequences. Any change causing alpine plants to cross the threshold from sub-zero temperatures (where there is minimal respiration) to one where temperatures are above zero even for a few days, will result in premature resumption of metabolic activity and a draw-down of metabolic reserves (Crawford, 2001). In terms of carbohydrate resource allocation, the phenological problem of optimising the timing of growth resumption in spring will be aggravated (Crawford, 2001).

Recent changes detected in the Swedish Scandes (Virtanen *et al.*, 2003; Kullman, 2004) may parallel future changes in the Highlands if seed banks of other species are present and in the absence of management intervention. Since the early 1950s, the range-margins of mountain birch (*Betula pubescens* ssp. *Tortuosa*), Norway spruce (*Picea abies*), Pinus sylvestris Scots pine (*Pinus sylvestris*), rowan (*Sorbus aucuparia*) and willows (*Salix spp*) have advanced by 120-375 metres to colonise moderate snow-bed communities (Kullman, 2002, 2004). Taken together, these findings indicate that these alpine ecosystems are already on the brink of a fundamental rearrangement and areal reduction (Kullman, 2006).

These changes have been in response to reduced summer snow-retention favouring seedling establishment and juvenile growth, with mild winters reducing the risk of frost desiccation and enhancing survival and height increment. In the Scottish uplands the distribution of snow and vegetation are closely linked, as topography and microclimate determine the available microhabitats. This dependence on late snow lie for thermal protection renders analogous communities in the Highlands particularly vulnerable to the direct combination of rising winter and spring temperatures. While changes to summer temperature and precipitation regimes are likely to directly affect the extent and duration of snowlie (and hence e.g. snowbed communities) and indirectly the mix of species as the century progresses.

6.3 Implications and limitations of the approach

As discussed previously in Sections 6.2.1 and 6.2.2 and as illustrated in Figures 6.1 and 6.2, modelled shifts in the isotherms associated with the upper limit of each vegetation zone are substantial when the selected UKCIP02 outputs are used to perturb the LRMs. Aside from the wider climate change uncertainties and the biases in the LRMs themselves previously discussed in Chapter 5, ecosystem responses to relatively rapid changes in temperature remain poorly understood.

For instance, for montane treelines, provided there is an accurate assessment of the lapse rate, the application of Koppen's rule that the treeline approximates to the 10°C T_{mean} isotherm for the warmest month in the year provides a basis for predicting the effect of temperature change (Koppen, 1931). Whereas in other work (at least for continental areas), mean summer temperature has been found to correlate with treeline position over large scales (Korner, 1999; Grace *et al.*, 2002; Moen *et al.*, 2004). Thus, while Figure 5.10 shows substantial altitidinal migration of mean summer T_{max} and T_{min} (and hence T_{mean}), it does not follow, nor is it implied, that vegetation zones will move in such a straightforward fashion as a direct linear response to temperature.

To underscore this, in continental areas with a lapse rate of 0.6°C/100 m, a 1°C incease in temperature would be expected to elevate the treeline by ~170m (Crawford, 2000). Whereas for higher oceanic-type lapse rate of 0.8°C/100 m, a 1°C rise in temperature would move the treeline upwards by ~125 m (Crawford, 2000). While several reports indicate an increased growth of established trees and shrubs (Kullman, 2003; 2004), as well as an increased shrub abundance (Sturm *et al.*, 2001) in response to recent regional warming.

However, the range of movement indicated above in response to recent warming has not been recorded so far (Crawford, 2000; Virtanen, 2003), although more recent studies suggest an acceleration of the trend in the upward shift of alpine plants since the mid-1980s (Walther *et al.*, 2005). In the case of continued warming, while some tree encroachment into the sub-alpine zone is likely, this is not likely to occur on a broad front since the altitudinal extent of treelines in the Highlands are strongly controlled by wind. Consequently local site topography and possible windfield changes are likely to remain a more significant local control under warming scenarios.

To emphasise the important control of wind on treeline in the Highlands, the LRM projections for the Onich and Balmoral 1961-1990 baseline values indicate the summer 10°C T_{mean} isotherm occurs around ~800 m and ~1000 m respectively. However, present treeline extents are ~300 m and ~600 m for western and eastern upland areas respectively. When primed with HadRM3 future time-slice outputs, the LRMs indicate migrations of the summer 10°C T_{mean} isotherm to altitudes of;

- ~1000 m and ~1200 m for the 2050s and 2080s for Onich baseline values;
- and ~1200 m and >1300 m (above the summits) for Balmoral baseline values.

But given the important limiting factor of wind in the present climate, it is highly unlikely that treeline will migrate to these elevations. Hence a more conspicuous change is likely to be the migration and expansion of alpine grasslands and dwarf-shrub heaths into higher altitudinal zones (Moen *et al.*, 2004; Kullman, 2006).

Since present confidence in GCM and RCM modelled windfield changes remain low (Hulme *et al.*, 2002), this interaction of temperature and windfield change in determining vegetation shifts is an area requiring further detailed modelling at the site scale. Certainly there are precedents for historically higher treelines, with paleoclimatic records indicating that treelines may have been substantially higher than at present during the boreal climatic optimum of the Holocene (Birks, 1988; Thompson and Brown, 1992; Grace *et al.*, 2002).

Monitoring studies elsewhere indicate a climate induced extirpation of high elevation floras and a northward shift in other groupings from the late-1980s (Lesica and McCune, 2004). While other work indicates that the montane pine ecoline appears to be stabilised by species interactions and not to be directly responsive to moderate climatic change (Hattenschwiller and Korner, 1995). However, other processes such as soil conditions, winter desiccation, seed limitation and competition are also important (Dullinger *et al.*, 2004). While yet other ecological modelling work has shown that plant species respond individually to migration pathways and local refugia (Grabherr, 2005).

It is clear that detailed multi-factorial and site-specific models are required, since changes driven by temperature, precipitation and windfield shifts will be compounded by other changes to the biotic and abiotic environment, e.g.:

- Rising CO_2 concentrations will enhance the growth of trees, although this may not be sustained on nutritionally poor sites, while longer growing seasons, higher temperatures and reduced cloud cover are also likely to promote tree growth. However, enhanced tree growth does not necessarily represent a positive impact on the wider woodland ecosystem. CO_2 fertilisation may lead to trees carrying more leaves and reduced light levels and water availability on the forest floor with implications for the competitive balance between species in the field layer.

- Other negative impacts include a greater risk of frost damage in spring and autumn, summer drought stress and an asynchrony between trophic levels within woodland ecosystems. High elevation woodland may suffer an increased risk of desiccation damage when soils are frozen and accompanied by warmer interludes (Section 1.3.2).
- Milder winters may result in an increased frequency or severity of existing pest and disease outbreaks, while mammalian species are likely to experience less winter mortality and increased immature survival with damage to young trees from grey squirrels and deer becoming more widespread.
- Negative impacts of higher winter rainfall include winter waterlogging, shallower root systems and hence increased wind damage. Reduced summer rainfall, by contrast, will lead to drought stress and more frequent fires.

Certainly the possible response to the more dramatic temperature changes across the altitudinal zones indicated by the RCM perturbed LRMs (Section 6.2), requires further work. Not least, since more detailed bio-climatic modelling work (Berry *et al.*, 2005) supports some of the broad findings (Section 6.2) and conclusions here. Specifically that there is a significant decrease in suitable climate space for the montane habitat, even under the 2050s scenarios (Berry *et al.*, 2005). However, detailed ecological modelling has indicated that the increase in the altitudinal range of some species is not likely to involve the simple elevational displacement of present vegetation zones (Gottfried *et al.*, 2002).

Overall therefore, the response of mountain plant species to climate change remains relatively uncertain due to a lack of accurate observational data on long-term changes and knowledge about adaptation (Klanderud and Birks, 2003; EEA, 2004). While our understanding of the potential impacts of climate change on ecosystem processes and biotic interactions remains limited (Ellis and McGowan, 2006). This recognition drives the requirement for data acquisition on the abundance and distribution of the most climate-vulnerable montane species, their reproductive abilities, dispersal capabilities and their general ecology, particularly in relation to their minimum ecosystem requirements (Ellis and Good, 2005). In addition, there is a need for developing methods which will deliver more locally relevant representations of future change to key climatic variables in order that land managers can make better informed decisions on stewardship of vulnerable communities and habitats. It is suggested that some of the approaches pursued here represent a positive step forward in this context.

Mountain research in general requires a truly inter-disciplinary and multi-disciplinary approach which encompasses the natural, health, social and engineering sciences (Diaz *et al.*, 2003), while of great current interest among both scientists and policymakers is the development of regional scenarios for future changes in climate extremes. In this context it is critical that the models realistically simulate all aspects of present-day climate, including climatological averages if unrealistic projected changes are to be avoided (Jones and Moberg, 2003). Findings here help underscore this conclusion, since technical problems associated with the simulation of baseline climate only add to the wider uncertainties surrounding future projections of altitudinal change. These strands of discussion are explored further in Chapters 7 and 8.

7. Summary and further discussion

Synopsis:

The major aims of this chapter are:

- To summarise the main findings of Chapters 1-5.
- To extend some discussion strands started in earlier Chapters.

7.1 Summary

Chapter 1 provides a framing context for much of the work presented in this thesis by providing a review of recent global and hemispheric trends, together with possible futures associated with assumptions of wider global warming. The Chapter also reviews aspects of important regional controls on NW European climate variability, as well as dealing with the possibility of climate 'surprises' (e.g. a THC shutdown). This in turn leads to a contextual framework for the altitudinal modelling component explored here in this climatically dynamic region. Finally, and given the narrative strand of 'uncertainty' permeating both this work and climate change projections in general, a brief review of sources of uncertainty in regional climate change scenarios is provided.

Chapter 2 explores a number of issues surrounding obtaining intact climatic data for temperature and precipitation on a sufficiently long baseline time-series to facilitate climate change impact assessments (CCIAs) for the region, including the evaluation of RCM/GCM baseline simulation outputs. Thereafter the various sections explore a number of methods aimed at using this data together with selective outputs from the HadRM3 RCM to arrive at more locally representative evaluations and projections of future changes to temperature and precipitation for the region.

The findings of the approach adopted, including how to quality control and apply the data are more fully explored in Chapters 3 and 4, and are further discussed in Sections 7.2.3 and 7.2.4 below. While the methods described in Section 2.5 utilising temperature lapse rate adjustment and corrections for orography are refined and explored further in Chapter 5 and the implications discussed in Section 7.2.4 below.

7.2 Further Discussion

7.2.1 Baseline construction and the decision not to grid datasets

Early in the project, a key decision was taken on whether or not to attempt gridding station values using some of the available spatial interpolation techniques. Various spatial methods for interpolating irregularly distributed station data to gridded long term averages (LTAs) have been tried. These include;

- polynomial regression (Goodale *et al.*, 1998);
- thin-plate splines (New *et al.*, 1999);
- kriging of regression residuals (Agnew and Palutikof, 2000);
- weighted local regression (Daly *et al.*, 2002) and inverse-distance weighting (Nalder and Wein, 1998).

Each have their own merits and obviously gridded data would have been more areally representative than station points for purposes of inter-comparing the HadRM3 outputs with observed data (an issue previously discussed in Section 4.2.5). This since it is generally accepted that grid box outputs from climate models have the spatial characteristics of areal averages (Osborn and Hulme, 1997; Frei *et al.*, 2003; Christensen *et al.*, 2004; Fowler *et al.*, 2005).

However, given the issues of data availability explored previously (e.g. a sparse station network in places) and the problem of extrapolating beyond station points, particularly for something as spatially variable as precipitation, the decision was taken not to grid datasets. In addition, recent work (Lloyd, 2005) has re-emphasised that with such a locally variable relationship as precipitation and elevation, there is no one best interpolation method.

This does not imply the approach is without worth if the time and computational resources are available. Indeed, such a grid-based approach has been recently developed for the UK (Perry and Hollis, 2005a, b) and used to examine trends and variability for a number of climatic variables across Scotland at a sub-regional resolution for the 1961-2005 period (Barnett *et al.*, 2006).

However, a weakness common to all gridding techniques is their inability to model localised effects, a problem which is exacerbated in complex, mountainous terrain. Thus, for example, even when advanced computational and statistical techniques are applied, station values may fit poorly when they are affected by local micro-climates, such as frost hollows (Perry and Hollis, 2005a).

A further complication is that the accuracy of grids will vary spatially according to both the density and representivity of the station network and the nature of the variable. Consequently the gridding exercise of Perry and Hollis (2005a, b) found the greatest errors were associated with the sparse station cover and complex terrain of the Highlands, with micro-climatic effects such as frost hollows creating a weakness in the gridding techniques. A further issue with the creation of these grids, certainly for the Highlands, was that Perry and Hollis (2005a, b) used a considerable number of station data with a relatively short intact record. Albeit they used a fairly sophisticated method to infill missing values (Section 7.2.2).

7.2.2 Procedures for infilling missing station data

As was discussed in Section 3.3, while correlations obtained from seasonal values would be preferable, an extension of the station correlation fits obtained from the annual values across the seasonal range are likely to be acceptable in the climatological context of this region. With a subdued seasonal variation of temperature common in the maritime climate (Pepin, 1995), there will be a corresponding suppression of the amplitude of variation in annual values. Hence the infilling exercise presented in Section 3.3 used only the station mean annual values in the correlation matrices. Nonetheless a useful future refinement to this aspect of the work would be to undertake this exercise using monthly and seasonal values and using the revised infilled station data values to repeat some of the experiments as discussed in Sections 3.3.1 and 3.3.2.

While the approach adopted for the station data infilling exercise was deemed the 'best practicable option' (Sections 3.3.1 and 3.3.2), there is clear scope to improve the rigour of the approach for future work, in order that the stations with the strongest and most reliable climatological relationship with the target station have the greatest input to the final estimate. Given the length of the baseline averaging period, a time series-based method could have been applied to station records and used for the infilling of missing values.

For example, the method applied by Tabony (1983) improves upon the constant difference approach by using linear regression against data from neighbouring stations with overlapping records. A modification of this approach was also adopted by Purves *et al.* (2000), while a further refinement of the method was applied by Perry and Hollis (2005a, b), whereby a regression-derived weighted average (based on the correlation coefficient) of estimates from six neighbour stations is used to arrive at the final estimate for the missing value.

For modelling precipitation, work applied in other regions (Kyriakidis *et al.*, 2001; Johansson and Chen, 2003) has included additional predictors in the regression models, such as wind and humidity, in order to allow for interaction with the lower atmosphere in precipitation mapping. While the problems associated with other statistical techniques, such as geographically weighted regression (GWR) (Brunsdon *et al.*, 2001) for this region have been broached previously in Section 4.3.

Despite these sort of issues surrounding the gridding of climatological data for the Highlands, an updated version of the gridded data produced by Perry and Hollis (2005 a, b) has been used recently to create a timely and useful regional analysis of trends and variations in a range of climatic variables by sub-region within Scotland (Barnett *et al.*, 2006) as has been briefly reviewed in Section 3.4.1.2. While Barnett *et al.* (2006) do not attribute an anthropogenic signature to observed trends in their analysis, they do note that many of the trends are significant and beyond the range of natural variability and often comparable with those predicted for the future. However, the reader is referred back to some of the cautionary discussion in Section 3.5 on cause and effect assignations.

7.2.3 Temperature simulation 1961-1990 – UKCIP02 HadRM3 outputs

In Chapter 4 the thrust of the approach was involved with establishing how station data may be used in order to evaluate RCM performance for selected variables in a data-sparse region such as the Highlands. Experimental design and the choice of the baseline period was predicated upon a supposition that, despite the considerable technical advance that HadRM3 represents by comparison with it's predecessors, parameterisation problems would remain in the representation of local climate for a region such as the Highlands. This initial working hypothesis tended to be substantiated by findings from work evaluating RCM performance in other European regions (e.g. Frei *et al.*, 2003; Christensen *et al.*, 2004; Deque *et al.*, 2005) as the project evolved.

Section 4.1.1 demonstrates that, despite the laudable attempts on the part of the modelling team at the Hadley Centre in differentially representing relief across the region in order to better resolve the spatial detail of climate. Paradoxically this creates an additional complication when trying to evaluate HadRM3 performance for the simulated 1961-1990 baseline when most available station data (at least for temperature) are at or close to sea level.

Consequently, methods of temperature lapse adjustment (Section 4.1.2) were developed and applied in order to better assess the performance of HadRM3 grid cells for the region. The improved correlations obtained reflected a reduced disparity between modelled and observed values following these adjustments and indicated the approach had some utility for the region, not least for the altitudinal projection of future changes (Chapter 5). The refinements arising from these findings have been explored further and applied both in a wider context (Section 6.1), and a selective case study context (Section 6.2).

Even with adjustment, however, the experimental approach flags parameterisation problems with the representation of certain seasonal values of T_{max} and T_{min} either within the HadRM3 grids or with the lateral boundary conditions of the driving AOGCM (see discussion at end of Section 7.2.5). However, due to the sizeable elevation corrections required via the LRMs for certain stations, some caution must be assigned to some of the values of ΔT°C reported here, an issue previously explored in Section 5.2.1 in the context of the more detailed altitudinal modelling of the Onich station data.

Despite doubts raised via the LRM corrections, monthly resolved data for selected stations and locations requiring little LRM correction were used to better elucidate some of the parameterisation problems associated with selected HadRM3 grids (Section 4.1.3). Issues relating to LRM performance are explored more fully in Chapters 5 and 6 and limitations of the approach discussed further in Section 7.2.6 below.

7.2.4 Precipitation simulation 1961-1990 – UKCIP02 HadRM3 outputs

Section 4.2 indicates that the spatial resolution of seasonal precipitation within HadRM3 remains problematic. However, the variability in precipitation simulated by climate models can be difficult to validate against station records because the level of variability in spatially averaged rainfall is not comparable to that of point rain-gauge observations (Mc Sweeney *et al.*, 2005), as was found to be the case here (Section 4.2.5). Compared to the representation of surface air temperature (SAT), this poorer ability to capture the local detail of precipitation patterns can be attributed to at least three factors:

- First, the dependence of rainfall on the driving GCM data at the ocean surface is lower;
- second, local rainfall variability is more chaotic than SAT (in both time and space), resulting in greater divergence between RCM realisations;
- last, its dependence on the RCM parameterisation schemes may be higher due to the greater complexity (and hence increased uncertainty) involved in modelling cloud and convective physics (Rowell, 2004).

However, these are issues that remain true for RCMs generally in topographically diverse regions as has been previously explored in Sections 4.2.5 and 4.3. The major inter-model differences in the precipitation frequency distributions, especially those in summer are primarily model intrinsic, rather than region dependant (Frei *et al.*, 2003). While the simulation of too dry conditions over the summer half-year in most of the RCMs also remains a prominent task for model improvement (Frei *et al.*, 2003). See also discussion at end of Section 7.2.5 below.

Even with the (relatively crude) corrections for orography undertaken for selected locations here (Section 4.2.5), the representation of precipitation appears to remain problematic for HadRM3. At least part of this is likely to be an artefact of the use of the intermediate-scale HadAM3H atmosphere only component in the model nesting procedures used to derive the UKCIP02 scenarios (Chapter 4 Introduction). Analogous problems have been encountered elsewhere, for example Bojario and Giorgi (2005) report an artificial shadowing effect for precipitation by comparison with observations for the central portion of southern Scandinavia, a feature they attributed to the relatively coarse representation of topography in HadAM3H.

In a UK context, these findings are not exclusive to the Highlands, since other work indicates HadRM3 performance is inconsistent across the UK, with no systematic error for any region or season (Mc Sweeney *et al.*, 2005). Consequently, projections of change in seasonal precipitation with elevation derived in this work (Section 5.5) remain conditional and subject to cautious interpretation, since they are based on outputs from only one RCM. Issues here are explored in more detail at the end of Section 7.5 and the wider implications are discussed in Chapter 8.

7.2.5 Remaining RCM limitations

This work and findings elsewhere (Frei *et al.*, 2003; Jones and Moberg, 2003) continue to indicate that substantial biases remain in the current generation of RCMs. For example, the under-representation of summer precipitation is well-documented in many RCMs for the European domain (Frei *et al.*, 2003; Chistensen *et al.*, 2004; Deque *et al.*, 2005). Therefore, while recent advances in climate system modelling have been substantial, findings here also emphasise the point that further advances are required before RCM outputs meet the local-scale projections required by policy makers (Schiermeier, 2004).

Some attempt has been made to broach this issue here, particularly for local projections of temperature change and, to a lesser extent, precipitation. While many aspects of the work are novel, as has been explored in preceding discussion, substantial difficulties remain. Not least since outputs from only one RCM with demonstrable systematic biases in the simulation of baseline climate have been used

(Sections 4.1 and 4.2). Overall, the results underscore the conclusions of Giorgi *et al.*, (2000, 2001) re-iterated by Jones and Moberg (2003) that further research aimed at reducing systematic errors in RCMs should be conducted.

In this respect, aspects of this work are both relevant and timely in that some of the locally reported findings here parallel weaknesses reported in RCM simulations for aspects of European regional climate elsewhere. For example, in a recent inter-RCM comparison over a European domain, a winter warm bias (Section 4.1) is also recorded as being particularly strong for Scandinavia (Christensen *et al.*, 2004). While the wider RCM problem with representation of summer precipitation Christensen *et al.* (2004) record is also documented here (Section 4.2). Christensen *et al.*, (2004) attribute this larger inter-model spread in summer to a failure on the part of the RCMs to capture the locally generated weather (e.g. convective rainfall) which is a particular feature of this season. Weaknesses here may be significant in terms of how accurately RCMs are capturing future patterns of summer precipitation change for eastern upland areas in the Highlands. Not least since high summer temperatures are known to trigger convectional rainfall events to the west of hills such as the Cairngorms.

This tends to suggest that the internal parameterisations of the RCMs are enhancing the dry and warm bias imposed by the boundary conditions of the driving AOGCM (Christensen *et al.*, 2004; Deque *et al.*, 2005). Whereas winter discrepancies suggest that the driving HadAM3H intermediate model is probably too zonal and simulating insufficient amounts of blocking events, the main source of inter-annual variability in this season (Christensen *et al.*, 2004). However, initial indications are that these representations, together with the representation of storm tracks, have been substantially improved in the new Hadley Centre Global Environmental Model (HadGEM1) (Ringer *et al.*, 2006). Overall in the European domain RCM inter-comparison, there is a warm bias with respect to observations in the extreme seasons, and a tendency to cold biases in the transition seasons (Christensen *et al.*, 2004; Deque *et al.*, 2005).

A final and obvious point in relation to climate change projections, and one previously made in Section 5.5 in relation to the orographic corrections for future precipitation changes. Is that if there are large scale atmospheric conditions in the predicted global climate that are beyond the range of empirical experience, then outputs from RCMs/GCMs with parameterisations conditioned upon present features of the climate system should be used with caution.

If it is accepted that climate is a complex chaotic system, the best any model can hope to do is produce a sequence of climate which, in terms of its properties is indistinguishable from the true trajectory within the period of interest. While climate models constitute an impressive human achievement, in terms of the detailed feedbacks within the earth system, they remain as relatively crude tools. Aside from many of the technical limitations of RCMs in adequately simulating the baseline climate explored and reported here and elsewhere, these sorts of parameterisation assumptions involving unchanged feedback mechanisms in a warmed atmosphere add to the wider uncertainties previously reviewed in Section 1.4.

7.2.6 Altitudinal modelling of temperature and corrections for orography

The application of the LRMs to the mean seasonal values take no account of important and fluctuating sub-seasonally resolved controls such as cloud cover, wind and snow cover which account for local differences in daytime T_{max} in particular at upland sites (Pepin and Norris, 2005; Pepin and Seidel, 2005). Late lying snow at high elevations in spring is a further potentially important control on lapse rates for upland areas in the Highlands since it cools surface air and hence steepen gradients (McClatchey, 1996). Other physical controls on ΔT such as the net radiation budget (Pepin and Seidel, 2005) were also (perforce) excluded from the seasonally-resolved LRMs.

Similarly, lapse rate values applied to seasonally resolved values cannot take account of the important synoptic controls associated with certain frontal types and the variation these introduce. Air mass characteristics are an important control on lapse rates, with synoptic types comprised of westerly components showing the most rapid decrease of temperature with elevation (Pepin, 2001). While McClatchey and Duane (1993) in the Cairngorms found steeper gradients in Polar or Arctic maritime air (Pm and Am) and Polar continental (Pc) air. By contrast shallower lapse rates were associated with Tropical maritime (Tm) or returned Polar maritime air (rPm). These steep temperature lapse rates are a particular feature of polar maritime air masses, especially in spring (Harding, 1978, 1979; Harrison, 1994; Pepin, 1995).

Lapse rates are also steeper when there is a strong altitudinal increase in cloud cover, decrease in sunshine duration, or strong wind shear (Pepin, 2001), with the fall in maximum temperature with height considerably more rapid than the fall in minimum temperature. McClatchey (1996) in the Cairngorms recorded temperature gradients of –7.0 to –10.0°C/km^{-1} during the middle of the day (1200-1500 GMT) when temperatures are at their highest. There is also considerable seasonal variation with the steepest lapse rates occurring in spring (particularly May) associated with the seasonal dominance of polar maritime air masses (McClatchey, 1996). Whereas, by contrast the shallowest lapse rates occur in November and December (Pepin, 1995; McClatchey, 1996).

Issues of wider uncertainty surrounding the use of scenario outputs from GCMs and RCMs have been discussed previously (Sections 1.4.1 and 1.4.2) and obviously these will be cascaded through the HadRM3 primed LRMs. While the performance of the HadRM3 outputs for the region have been evaluated seasonally (Section 4.1.2) and the HadRM3 warm and cold biases for selected locations more fully explored at a monthly resolution (Section 4.1.3).

Similarly, the cold biases of the LRMs (at least for higher elevations) have been explored in Sections 5.1.2 and 5.1.3. Despite the wider uncertainties associated with applying the RCM climate change outputs and some of the bias problems with the LRMs, it is suggested that work undertaken here represents a considerable advance in terms of delivering locally relevant projections of possible temperature change with altitude (see e.g. Moen *et al.*, 2004).

Nonetheless, further qualifications remain surrounding the approach. The approach adopted involves inherent assumptions, modelling changes to the microclimatic growth environment for example were beyond the scope of time and resources available to this project (see further discussion in Chapter 8). Similarly, potential changes to inter-specific competition as lowland species migrate upslope cannot be taken into account. This highlights the overall difficulty of predicting the effect of climatic warming on mountain flora and vegetation due to the complexity of the interactions and feedbacks operating within the whole ecosystem (Klanderud and Birks, 2003). In addition, the response of mountain plant species to climate change also remains relatively uncertain due to a lack of accurate observational data on long-term changes and knowledge about adaptation (Klanderud and Birks, 2003; EEA, 2004).

These sorts of difficulties emphasise the highly inter-disciplinary research effort required for conducting valid and locally representative CCIAs in our upland environments. Consequently, discussion in subsequent sections here takes place against this backdrop of remaining technical problems associated with obtaining valid climate change projections which can be applied to higher elevation sites, together with the recognition that a highly inter-disciplinary research effort is an absolute requirement. Issues here are also explored further in Chapter 8.

8. Conclusions and Suggestions for Future Research

8.1 Conclusions

In summary this work has shown that:

(i) Despite the problem of climate data sparsity for much of the Highlands, with appropriate quality control procedures (Chapter 3) and temperature lapse rate adjustment, some measure of RCM performance evaluation can be undertaken for the region (Chapter 4).

(ii) There is substantial intra-region variability in temperature and precipitation data over the baseline period. However, winter-time temperature anomalies and precipitation total fluctuations for western stations in particular have demonstrated linkages to the behaviour of the winter NAO when smoothed appropriately (Chapter 3).

(iii) Variably combining station and RCM output data for selected climate change scenarios at representative locations allows for more refined local-scale projections of possible isotherm shifts than has hitherto been available (Chapters 5 and 6). Although only applied to selected isotherms and case studies in this work (Chapters 5 and 6), it is suggested that the approach has considerable wider utility for conducting more comprehensive local-scale CCIAs for upland sites.

This work also demonstrates that remaining challenges include:

(i) Scale-dependant controls on local topo-climates here are not adequately captured in the present generation of GCMs/RCMs (Chapter 4). Consequently, until technical advances in modelling improve the representation of localised climate processes in topographically diverse regions such as the Highlands, it is incumbent upon the research community to develop methods to better represent localised projections of future change.

(ii) Observed data sparsity in this region (particularly for higher elevation sites) is an issue hampering the performance assessment and ground truthing of outputs from this and subsequent generations of RCMs (Chapters 4 and 5). This flags a clear need for an expanded and more spatially and altitudinally representative upland station network in order that future generations of climate researchers can better monitor ongoing climatic change as the century progresses. Such a network would also allow the modelling community to provide more locally robust projections of change which would better inform an evolving future policy response to ongoing climatic changes.

(iii) Despite these sorts of issues, deploying locally representative corrections for topography and orography allows for improved local-scale projections of possible future changes to key climatic variables using data presently available (Chapters 4 and 5). In turn, these can be used to undertake more locally resolved CCIAs in a case study context (Chapter 6). However, residual uncertainties associated with the use of model outputs remain.

In addition, the work has contributed to the understanding of:

(i) The importance of localised site controls in determining climate variability across the Highlands (Chapter 3).

(ii) How baseline data on an appropriate record length may be obtained and quality controlled for purposes of evaluating climate model outputs (Chapters 2 and 3).

(iii) Aspects of variability in regional data and linkages to wider hemispheric controls, e.g. the influence of the NAO.

(iv) How baseline data may be variably combined with RCM outputs and LRM models to project future changes to different elevations for selected locations, and how outputs derived here may then be used to undertake selective upland CCIAs (Chapters 5 and 6).

(v) The work has also helped to underscore the inherent difficulty and uncertainty associated with obtaining localised projections of climate change.

8.1.1 Wider conclusions and synthesis – regional climate change assessment

With the next set of UKCIP (UKCIP-Next) scenarios now scheduled for spring 2008 (Chris West, pers. comm.), and with plans to run the next Hadley Centre RCM using boundary conditions from the new Hadley Centre AOCGCM (HadGEM1; Johns *et al.*, 2004; Johns *et al.*, 2006; Martin *et al.*, 2006; Ringer *et al.*, 2006; Stott *et al.*, 2006) at a 25km x 25km resolution to generate these (Geoff Jenkins, pers. comm), there is clear scope to extend some of the approaches explored here. However, it remains unlikely that the next or even subsequent generation of GCMs or RCMs will provide the required predictive skill level required for site specific CCIAs at upland locations. This is not a criticism of GCMs, despite the remaining problems, they undoubtedly remain the most comprehensive tool for studying climate change (Raisanen and Doscher, 1999; Lange, 2001; Christensen *et al.*, 2004; Wang *et al.*, 2004). Rather it is unreasonable to expect climate system modelling at the spatial resolution provided by GCMs and RCMs to capture an (upland) micro-climatic variation which can be measured in metres.

Micro-topographic variation is expressed through the relative degrees of shelter and exposure provided by any particular site, and operates mainly through the microclimatic growth environment of the individual plant. Apart from the generalised effect of micro-topography in terms of shelter and exposure, specific topographic features form particular and specialised habitats for plant growth. Perhaps the most important are the ledges and fissures of mountain cliffs, block-fields, block-scree, scree-slopes and gravel slides, patterned ground and solifluction features together with drainage and seepage lines - the wet flush habitats. All of these relatively small scale features tend to interact with microclimate, but their peculiarity as plant habitats is also associated with their inherent instability and dynamic nature, together with their effects on soil moisture and soil nutrient status.

Given the close interplay between the abiotic and biotic environment at these locations, it is abundantly clear that methods have to be developed which can undertake CCIAs at a micro-climatic resolution. As some of the approaches explored here illustrate, more site-specific local projections can be obtained, offering the possibility of developing site and CCIA-specific local scenarios of change. It is suggested that with refinement and extension (see below), some of the methods developed and explored here could represent a departure point for more detailed modelling at an improved spatial resolution.

There are wider issues surrounding the validity of taking areally averaged data supplied by the dynamically downscaled output an RCM provides and applying this to station points as has been the case here. However, these are less substantial for the more spatially coherent variable of temperature than they are for precipitation.

Nonetheless, and as has been selectively explored in Chapters 5 and 6, by variably combining such data and by providing a local component of altitudinal and orographic correction, a wider range of possible future changes can be projected. Although as has been explored throughout this work, any approach using data outputs from GCMs cannot constrain the wider uncertainties inherent to projections of future change to climatic variables (Section 1.4).

The UKCIP-Next scenarios intend to try and constrain uncertainty by providing the impacts community with a probability assignation surrounding future projections of change from the next Hadley Centre model (HadGEM1), as well as from a number of other GCMs (Murphy *et al.*, 2004; Stainforth *et al.*, 2005; Geoff Jenkins pers. comm.). However, despite these advances offered by the modelling community, there are remaining concerns surrounding the validity of the socio-economic and development trajectory scenario assumptions used to derive the GHG emissions scenarios driving the models.

With the IPCC having decided there was no time to develop new scenarios for its 4th assessment report, the various modelling groups are using the same SRES scenarios (Table 2.1) used in the previous assessment (Schiermeier, 2006). This has led to criticism from many economists on the basis that the existing SRES scenarios rely on outdated economic theories which fail to reflect how lifestyle and energy demand in both rich and poor countries are likely to change (Schiermeier, 2006; Van Vuuren and O'Neill, 2006). Therefore, despite the improved probabilistic assessments UKCIP-Next intend to provide, these permeating uncertainties surrounding the SRES scenarios will remain (Geoff Jenkins, pers. comm.).

There is consensus that irrespective of present mitigation strategies, the current level of GHGs already commits us to some level of warming over the coming decades. Globally averaged concentrations of carbon dioxide (CO_2), methane (CH_4) and nitrous oxide (N_2O) in the Earth's atmosphere reached their highest-ever recorded levels in 2004, CO_2 was recorded at 377.1 parts per million (ppm), CH_4 at 1783 parts per billion (ppb), and N_2O at 318.6 ppb (WMO, 2006). These values supersede those of pre-industrial times by 35%, 155% and 18% respectively, an increase over the previous decade by 19ppm, 37ppb and 8ppb in absolute amounts (WMO, 2006). However, while much has been made recently of 'tipping point' exceedance in a climate context precipitating rapid change beyond a notional threshold in both the scientific literature and the media, in a balanced view, the human tipping points are likely to be more important, with a plethora of development choices and mitigation options available. Therefore while the inertia of the climate system commits us to some component of anthropogenically induced warming over the next few decades, present human activities have not yet defined the magnitude of later century warming. As a result development pathway choices over the next couple of decades are more likely to influence the possible magnitude of end of century warming.

Accordingly, and in recognition of the importance of policy driven mitigation initiatives, the substitution of fossil fuels with the increasing use of renewable energy sources is now recognised by the UK Government as a fundamental priority in reducing GHG emissions. Consequently research into new renewable energy technology and initiatives is proceeding at a frantic pace in order to mitigate against the likely impacts of climatic warming, as well as to address wider strategic issues related to finite reserves of carbon-based fuels and the future security of energy supplies. However, it is likely that, even with ongoing technology development, assumptions about the future socio-economic and technological trajectories underpinning the SRES scenarios e.g. are likely to remain one of the most sizeable sources of uncertainty in projections of climate change.

It is obvious that there is little that local research and impact assessment communities can do in attempting to reconcile these wider issues in terms of trying to derive localised projections which better inform policy. However, if we accept the need for a focussed and highly-interdisciplinary research effort, both for purposes of conducting CCIAs in the uplands and for more locally relevant projections of change generally. Some of the methods adopted here should help inform an improved approach in the future.

It is also apparent from elements of preceding discussion that there is no one 'best' RCM or GCM in terms of how they perform technically in topographically diverse regions (Frei *et al.*, 2003; Christensen *et al.*, 2004). However, elements of this work demonstrate that combining outputs with station data and incorporating some element of altitudinal projection can offer added utility. At least insofar as providing a wider range of numbers on the possible magnitude of future change until more refined GCMs/RCMs primed with realistic development scenarios become available. In terms of constructing site-specific CCIAs which best span the range of wider uncertainties, the most obvious extension to this work would be to take the outputs available from the full range of RCMs and GCMs and apply these locally in order to envelope the full range of projected future changes. Inherent in this is the idea that a full array of GCM/RCM outputs could also be applied to observed baseline data, as well as used directly, and with some component of altitudinal correction applied.

It is also clear from much of this work that despite the advances in climate system modelling that we are still some distance away from the local-scale projections of future change required to inform a robust policy response. This work has helped to demonstrate that substantial challenges remain in obtaining valid projections of future change for topographically diverse regions, a situation which is likely to remain true in the foreseeable future for mountain regions generally.

Some clear limitations of the approaches attempted here (aside from the use of output data from only one RCM) include;

- the spatial resolution of projected changes (Chapter 5), including issues relating to the application of areally averaged (RCM) data to station points;
- the limited temporal resolution of possible future changes offered by using only mean seasonal values.

8.2 Suggestions for Future Research

The obvious solution to the first point above would be to apply a suite of RCM/GCM outputs to the 5km x 5km gridded climatology (Perry and Hollis, 2005a, b) available via the Met Office and UKCIP, and to apply LRM and orography corrections. This would offer the advantage of extending the approach spatially in the Highlands, particularly to regions identified as data sparse (e.g. Sutherland and Caithness) in this work. However, some of the issues surrounding interpolation explored in Section 7.2.1 would apply to the use of such data. Indeed, Perry and Hollis (2005a, b) provide such a caveat themselves, while Barnett *et al.* (2006) re-iterate this cautionary note on the use of interpolated data in the Highlands.

Whereas the second point on the temporal resolution of climate change data and how it may be applied for purposes of local CCIAs constitutes a more substantial challenge for the research community. If it is accepted that the reductionist paradigm of classical science is not best suited to addressing the complexity inherent in conducting CCIAs, but rather a systems-based approach linked across disciplines, implicit in this is the notion that research questions should be clearly and specifically directed in order to focus the methods and approach on a specified outcome. Ideally this would follow stakeholder consultation to frame the objectives prior to undertaking a focussed research effort which engages across disciplines. In this sense, an iterative approach which is both reflexive and outcome centred is envisaged as being the most appropriate way to proceed.

The approach adopted in aspects of this work provides a relatively crude temporal (seasonal mean) projection of change only. Whereas the ecological modelling community, for example, are likely to require data inputs at a daily or sub-daily resolution and spanning changes to an array of climatic variables (e.g. accumulated degree days), together with an indication of extreme event frequency changes. Data at this temporal resolution is available from many of the climate models, and hence multiple or ensemble outputs from a number of models could be projected with altitudinal corrections to prime ecological models for both site and purpose-specific CCIAs.

However, the scale of the datasets involved would require a substantial input from the data handling and programming community to realise such projections. As has been discussed throughout, this would not resolve the inherent uncertainties associated with using climate model projections. Nonetheless, such an approach would provide a better resolved range of possible changes, together with a more realistic local-scale scenario-based assessment tool both for purposes of land management contingency assessment and policy planning.

In summary form, this work has helped to identify that further work in some of the following areas is required:

(i) Arriving at an improved temporal resolution for projected local/upland changes using outputs from an array of GCMs/RCMs realised at a daily or sub-daily scale. Projections at this scale would better envelope the range of uncertainties associated with climate model outputs at a localised scale, particularly since outputs from subsequent generations of GCMs will provide probability assignations associated with projections of change for key climatic variables.

(ii) The development of integrated localised assessment models (climatic, ecological and socio-economic) linked to a trans-disciplinary scoping and research effort. Implicit in this is the notion of fully joined up and effective communication between local, national and international research groupings across disciplines, together with a full stakeholder engagement.

(iii) To somehow integrate outputs from a wider intellectual effort seeking to resolve scale dependent issues in order to better inform both mitigation and adaptation strategies in relation to anticipated climate change. Aside from equity and justice issues (both social and generational) in relation to the differential impacts of projected climatic changes, how are 'sustainable' local communities delineated and defined when their own efforts are increasingly subsumed in a global business as usual trajectory, particularly as the future economic powerhouses of India and China e.g. come fully 'on stream' in GHG terms? In the sense of a scale-dependent mis-match between the likely future drivers of anthropogenic interference in the climate system (global) and the impacts (local) of a changing climate; i.e. in terms of changing source areas of future GHG emissions in a differentially industrialised future society, and the points of impact of future weather events in a changed climate, how can a 'normalised' model of what constitutes local sustainability be defined and applied when more than a climatic threshold has been exceeded in a future society?

Appendix Figures and Tables

Chapter 2

Table A2.1: HadRM3 Output Variables (Source: Hulme *et al.*, 2002)

Variable	Code	Defintion
Maximum temperature (°C)	TMAX	
Minimum temperature (°C)	TMIN	
Daily mean temperature (°C)	TEMP	1.5m temperature averaged over the daily cycle (i.e. over all the five minute timesteps)
Total precipitation rate (mm/day)	PREC	
Snowfall rate (mm/day)	SNOW	Rate of accumulation of equivalent water (ie. all snow melted)
10 m wind speed (m/s)	WIND	
Relative humidity (%)	RHUM	
Total cloud in longwave radiation (fraction)	TCLW	Fractional cloud cover
Net surface longwave flux (W m^{-2})	NSLW	$L_{up} - L_{down}$ (usually negative, ie. upward)
Net surface shortwave flux (W m^{-2})	NSSW	DSWF multiplied by (1 – ground albedo)
Total downward surface shortwave flux (W m^{-2})	DSWF	direct + diffuse solar radiation
Soil moisture content (mm)	SMOI	moisture in the root zone (ie. available for evapotranspiration)
Mean sea level pressure (mb)	MSLP	
Surface latent heat flux (W m^{-2})	SLHF	energy lost by the surface due to evaporation of water from the soil or the vegetation canopy.
Specific humidity (g/kg)	SPHU	

Table A2.2: Spatial Reference and Elevation Information For HadRM3 'Sea' Boxes

Grid Box Number	Latitude	Longitude	Elevation
73	58.89	-5.48	0
74	58.98	-4.65	0
75	59.06	-3.82	0
76	59.14	-2.99	0
90	58.36	-6.11	0
91	58.46	-5.30	0
94	58.71	-2.84	0
106	57.84	-6.73	0
107	57.93	-5.93	0
111	58.28	-2.69	0
123	57.41	-6.55	0
127	57.77	-3.35	0
128	57.84	-2.54	0
129	57.92	-1.73	0
141	56.98	-6.36	0
147	57.48	-1.60	0
159	56.55	-6.18	0
164	56.98	-2.27	0
165	57.05	-1.47	0
177	56.12	-6.01	0
182	56.54	-2.13	0

Chapter 3

Table A3.1: 1961-1990 baseline annual and seasonal values, temperature station mean maxima (T_{max}) °C

ID Number	Station Name	Annual T_{max}	Spring T_{max}	Summer T_{max}	Autumn T_{max}	Winter T_{max}
1	Aberdeen	11.05	9.82	16.51	11.77	5.96
2	Balmoral	10.38	9.34	16.86	10.75	4.39
3	Craibstone	10.90	9.74	16.54	11.62	5.54
4	Fyvie Castle	11.33	10.43	17.36	11.67	5.49
5	Arbroath	11.25	9.97	16.72	12.04	6.13
6	Dundee	11.91	10.98	18.10	12.45	5.85
7	Montrose	11.19	10.10	16.95	11.88	5.67
8	Mylnefield	11.75	10.76	17.89	12.28	5.85
9	Fort Augustus	11.52	10.85	17.43	11.70	5.83
10	Inverness	11.94	11.11	17.54	12.34	6.57
11	Isle of Rhum	11.82	10.99	16.62	12.23	7.25
12	Onich	12.13	11.58	17.53	12.39	6.82
13	Prabost	11.00	10.37	15.94	11.38	6.10
14	Rannoch School	10.90	10.40	17.31	11.02	4.64
15	Faskally	11.72	11.12	18.11	11.95	5.48
16	Fortrose	11.05	9.79	16.21	11.72	6.34
17	Kinlochewe	12.05	11.49	17.60	12.23	6.64
18	Poolewe	11.82	11.07	16.73	12.23	7.23
19	Cape Wrath	8.98	7.39	12.50	10.04	5.87
20	Tiree	11.30	10.33	15.50	12.17	7.50
21	Ardtalnaig	11.50	11.13	18.13	11.50	5.53
22	Braemar	10.10	9.33	16.87	7.63	4.10
23	Kinloss	11.80	11.03	17.60	9.77	6.53
24	Stornoway	10.80	9.87	15.20	9.33	6.97
25	Kirkwall	10.20	9.10	14.80	9.03	6.03

Table A3.2: 1961-1990 baseline annual and seasonal values, temperature station mean minima (T_{min}) °C

ID Number	Station Name	Annual T_{min}	Spring T_{min}	Summer T_{min}	Autumn T_{min}	Winter T_{min}
1	Aberdeen	4.79	3.29	9.61	5.50	0.69
2	Balmoral	2.34	0.96	7.19	3.15	-2.01
3	Craibstone	4.49	2.93	8.99	5.43	0.52
4	Fyvie Castle	3.56	2.38	8.64	4.02	-1.03
5	Arbroath	5.57	4.07	10.15	6.51	1.48
6	Dundee	5.41	4.10	10.22	6.17	0.99
7	Montrose	4.91	3.48	9.54	5.69	0.84
8	Mylnefield	4.97	3.72	9.80	5.71	0.53
9	Fort Augustus	4.75	3.29	9.76	5.47	0.26
10	Inverness	5.36	4.03	10.37	6.04	0.92
11	Isle of Rhum	5.62	4.00	9.61	6.63	2.20
12	Onich	4.89	3.42	9.29	5.73	1.03
13	Prabost	5.12	3.67	9.32	6.03	1.35
14	Rannoch School	2.62	1.03	7.49	3.48	-1.69
15	Faskally	3.97	2.61	9.23	4.61	-0.66
16	Fortrose	5.99	4.38	10.81	6.84	1.82
17	Kinlochewe	4.66	3.27	9.31	5.37	0.54
18	Poolewe	5.81	4.28	10.20	6.70	2.10
19	Cape Wrath	5.91	4.24	9.50	7.01	2.80
20	Tiree	6.40	4.80	8.87	7.57	3.10
21	Ardtalnaig	4.60	3.13	9.37	5.47	0.60
22	Braemar	2.50	1.10	7.63	3.40	-1.97
23	Kinloss	4.70	5.77	9.77	5.57	0.30
24	Stornoway	5.20	3.83	9.33	5.97	1.80
25	Kirkwall	5.00	3.47	9.03	4.07	1.67

Table A3.3: 1961-1990 baseline annual and seasonal values; temperature station mean maxima (T_{max}) °C for stations where infilled values applied

ID Number	Station Name	Annual T_{max}	Spring T_{max}	Summer T_{max}	Autumn T_{max}	Winter T_{max}
1	Aberdeen	11.02	9.81	16.45	11.79	5.87
4	Fyvie Castle	11.21	10.29	17.10	11.65	5.49
6	Dundee	11.91	10.93	18.09	12.43	5.99
9	Fort Augustus	11.38	10.65	17.29	11.60	5.68
10	Inverness	11.80	10.92	17.51	12.23	6.36
11	Isle of Rhum	11.80	10.96	16.68	12.27	7.07
12	Onich	12.08	11.52	17.43	12.37	6.83
13	Prabost	11.10	10.44	16.02	11.44	6.29
14	Rannoch School	11.11	10.51	17.40	11.29	5.14
16	Fortrose	11.14	9.96	16.39	11.88	6.12
17	Kinlochewe	11.98	11.39	17.43	12.24	6.64
18	Poolewe	11.73	10.97	16.59	12.16	7.00

Table A3.4: 1961-1990 baseline annual and seasonal values; temperature station mean minima (T_{min}) °C for stations where infilled values applied

ID Number	Station	Annual T_{min}	Spring T_{min}	Summer T_{min}	Autumn T_{min}	Winter T_{min}
1	Aberdeen	4.76	3.26	9.55	5.53	0.61
4	Fyvie Castle	3.26	2.05	8.28	3.77	-1.26
6	Dundee	5.46	4.13	10.26	6.22	1.14
9	Fort Augustus	4.85	3.40	9.80	5.58	0.42
10	Inverness	5.37	4.03	10.37	6.07	0.92
11	Isle of Rhum	5.60	4.02	9.63	6.63	2.05
12	Onich	4.92	3.48	9.28	5.75	1.13
13	Prabost	5.16	3.69	9.34	6.04	1.46
14	Rannoch School	3.17	1.63	7.96	4.01	-0.79
16	Fortrose	5.73	4.17	10.45	6.63	1.76
17	Kinlochewe	4.62	3.20	9.42	5.39	0.36
18	Poolewe	5.76	4.22	10.10	6.66	1.96

Table A3.5: 1961-1990 baseline annual and seasonal values, precipitation (Precip) station totals (mm)

ID Number	Station Name	Annual Precip	Spring Precip	Summer Precip	Autumn Precip	Winter Precip
1	Lumphanan	810.71	170.44	181.73	213.31	242.64
2	Balnakeilly	914.08	192.50	195.54	257.20	269.68
3	Barney	1116.68	248.32	226.88	303.39	338.03
4	Clintlaw	1089.89	240.10	218.99	298.54	332.72
5	Glen Lee	1417.08	331.17	263.62	363.55	458.51
6	Glenhead	1162.80	255.91	232.89	319.76	354.26
7	Longdrum	1098.75	248.06	217.85	297.07	336.26
8	Newton	992.35	211.66	206.21	273.81	293.66
9	The Linns	1250.03	285.07	235.93	333.17	400.32
10	Westerton	1061.38	237.19	212.90	289.78	321.47
11	Allt Leachdach	1762.57	394.34	268.82	477.43	623.55
12	Alltbeithe	3052.12	690.30	467.21	831.17	1064.55
13	Fort William	2088.78	466.78	317.36	584.43	718.81
14	Glen Einich	1331.25	290.18	275.70	361.29	398.01
15	Glenfeshie Lodge	1004.63	217.28	190.58	266.30	330.84
16	Loch a' Chrathaich	1561.11	354.36	255.71	409.00	538.93
17	Loch Eilde Mhor	2795.75	605.45	449.26	517.09	939.76
18	Loch Pattack	1678.07	366.50	269.40	473.11	570.14
19	Lochan na Sgud	2690.30	584.08	400.75	749.04	956.72
20	Lon Mor	1445.01	325.34	231.54	393.54	503.28
21	Rhum, Coire Dubh	3227.75	608.24	613.53	1016.52	991.94
22	Rhum, Harris	1564.22	314.50	278.54	465.61	485.44
23	Rhum, Kilmory	1944.32	402.71	312.13	571.50	657.94
24	Rhum, Slugan Burn	3240.32	639.66	560.12	956.47	1050.08
25	Skye, Alldearg House	3218.90	668.22	515.64	933.25	1104.35
26	Sronlairig Lodge	1458.56	331.29	240.20	386.10	504.67
27	Tromie Dam	1361.47	340.28	219.91	340.30	471.13
28	Allt Dubhaig	1505.63	351.07	241.85	401.16	494.49

Table A3.5 (contd): 1961-1990 baseline annual and seasonal values, precipitation (Precip) station totals (mm)

ID Number	Station	Annual Precip	Spring Precip	Summer Precip	Autumn Precip	Winter Precip
29	Allt Girnaig	907.46	362.09	188.49	253.53	287.61
30	Allt na Bogair	1483.59	348.25	236.62	379.22	517.84
31	Ben Vorlich	2187.30	494.04	372.25	595.08	705.58
32	Bruar Intake	1153.81	251.02	204.37	303.99	397.54
33	Capel Hill North	1023.28	213.81	219.44	274.93	338.49
34	Capel Hill West	981.40	202.14	210.97	266.17	302.63
35	Chapel Burn	1182.76	250.17	240.77	350.60	372.86
36	Frandy Burn	1742.71	399.33	312.60	468.20	565.55
37	Loch Benally South	948.82	190.81	207.97	267.92	282.78
38	Loch Ordie	942.87	196.32	206.44	256.64	291.63
39	Lochan Oissineach Mhor	1002.63	208.29	211.01	271.70	312.45
40	Slateford Burn	1117.86	249.66	224.50	296.39	337.40
41	Sronphadruig Lodge	1638.83	373.81	284.00	432.86	551.80
42	Am Meallan	1861.52	402.79	317.48	536.37	605.73
43	Carn Anthony	2881.40	640.42	467.49	806.05	916.03
44	Coulin Pass	2590.72	606.51	409.34	722.62	866.78
45	Gleann Fhiodhaig	2093.13	506.17	314.52	549.07	696.18
46	Glenshiel Forest	3463.86	917.73	481.96	816.17	1273.51
47	Grudie Bridge	2150.86	480.19	297.48	610.26	778.09
48	Heights of Kinlochewe	1987.26	428.31	300.06	553.54	689.59
49	Loch Fannich South	2221.16	513.61	349.47	582.06	778.95
50	Loch Glass	1287.72	297.46	208.93	329.47	454.20
51	Loch Gleann na Muice	2460.23	532.49	389.39	679.73	847.99
52	Luibachlaggan	1696.54	412.81	249.78	424.45	608.54
53	Strathrannoch	1317.90	301.38	213.61	338.38	462.78
54	Strone Nea	1734.16	474.69	260.74	394.63	589.44
55	Tornapress	2270.34	496.71	361.22	662.93	748.69

Table A3.6: 1961-1990 baseline annual and seasonal values, precipitation (Precip) station totals (mm) following infilling

ID Number	Station Name	Annual Precip	Spring Precip	Summer Precip	Autumn Precip	Winter Precip
1	Lumphanan	808.00	169.18	183.33	215.20	241.39
12	Alltbeithe	2992.72	675.92	462.66	816.74	1036.75
13	Fort William	1996.65	437.21	319.43	564.19	677.10
14	Glen Einich	1420.44	307.73	281.12	386.68	448.36
15	Glenfeshie Lodge	1052.87	230.53	204.06	276.18	336.84
16	Loch a' Chrathaich	1577.13	353.94	261.90	419.19	545.76
20	Lon Mor	1545.57	339.52	241.50	434.64	528.61
22	Rhum, Harris	1601.47	319.36	291.79	482.33	507.86
26	Sronlairig Lodge	1459.19	325.51	244.02	387.71	508.04
27	Tromie Dam	1318.26	320.55	225.86	337.59	444.38
28	Allt Dubhaig	1500.72	345.10	245.95	407.48	505.05
29	Allt Girnaig	916.17	353.50	189.63	255.71	283.90
30	Allt na Bogair	1467.65	343.24	242.10	383.12	499.20
31	Ben Vorlich	2138.16	483.09	364.67	581.61	711.93
32	Bruar Intake	1137.43	247.29	204.43	302.89	383.99
40	Slateford Burn	1119.90	246.52	227.18	298.17	349.71
43	Carn Anthony	2670.34	587.18	436.07	748.68	838.26
44	Coulin Pass	2480.51	562.44	401.19	698.67	828.93
45	Gleann Fhiodhaig	2033.63	487.99	307.38	532.13	705.23
46	Glenshiel Forest	3331.89	821.18	519.83	844.99	1162.82
47	Grudie Bridge	2370.70	540.00	360.45	644.83	816.77
48	Heights of Kinlochewe	1984.10	442.70	295.57	548.38	691.41
49	Loch Fannich South	2195.88	503.69	340.87	576.88	772.29
51	Loch Gleann na Muice	2502.87	548.89	392.64	680.13	855.84
52	Luibachlaggan	1711.44	408.76	258.82	434.39	611.44
54	Strone Nea	1736.47	461.77	268.75	409.24	583.42
55	Tornapress	2164.82	475.28	352.01	629.58	703.49

Chapter 6

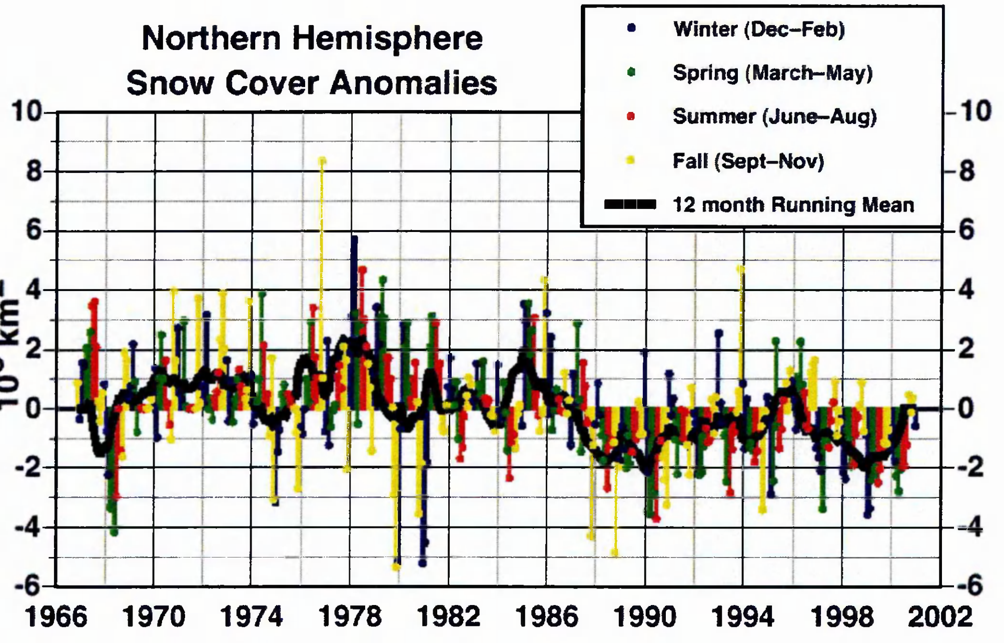


Figure A6.2: Snow cover anomalies in the Northern Hemisphere from the weekly National Oceanic and Atmospheric Administration (NOAA) dataset (Source: Morison *et al.*, 2001)

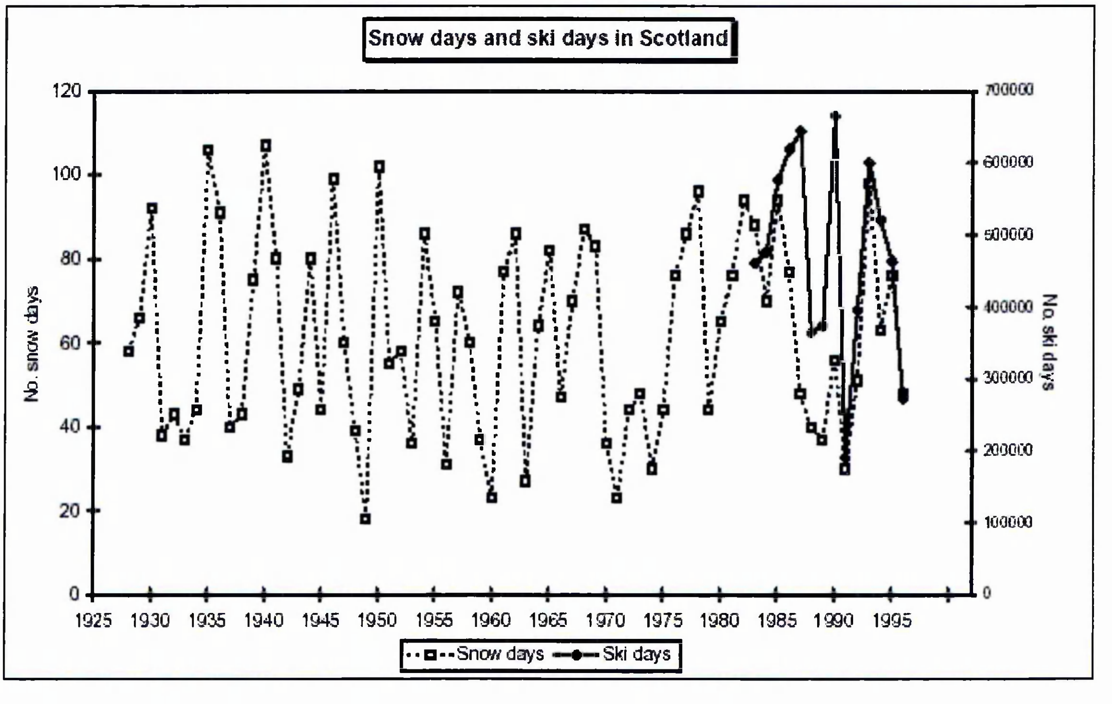


Figure A6.3: Number of days with lying snow each year at Braemar (dashed line) since 1927, and the more recent record of ski-days at the five main Scottish skiing centres (solid line), (Source: Palutikof, 1999).

Appendice Figures – Landscape Layout

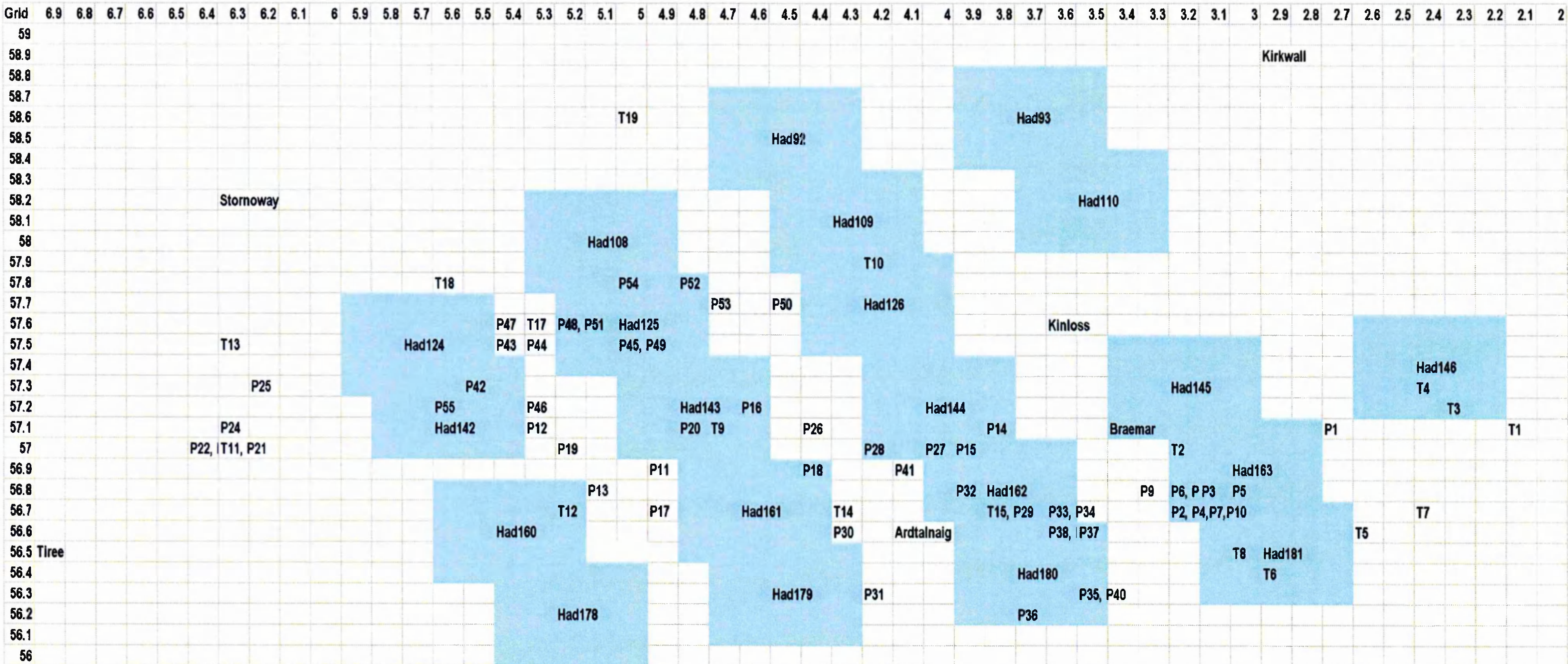


Figure A2.1: Illustration of spreadsheet matrix used to reference stations to the corresponding HadRM3 land grid cell. Longitude and latitude co-ordinates are on the x and y-axes respectively. HadRM3 grids are shaded and labelled by number. Temperature and precipitation stations are prefixed by T and P and numerically coded. Met Office temperature station records added to improve spatial cover are identified by name.

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Publications and Related Outputs

Copies of publications and exemplar peer-reviewed conference contributions are provided in chronological order on the attached CD-ROM, these are;

1. **Sensitivity of ferry services to the Western Isles of Scotland to changes in wave climate.**
David Woolf, **John Coll**, Stuart Gibb and Peter Challenor. (2004). Paper presented at OMAE conference, Vancouver, Canada.

2. **Modelling future climate in the Scottish Highlands - an approach integrating local climatic variables and climate model outputs**
John Coll, Stuart W. Gibb and John Harrison. (2005). In *The Mountains Of Europe: Conservation, Management, People and Nature*. Eds. D.B.A. Thompson, M.F. Price and C.A. Galbraith. HMSO, Edinburgh. pp 103-119.

3. **Sensitivity of the Western Isles of Scotland to changes in wave and wind climate.** David Woolf, **John Coll**, Stuart Gibb and Peter Challenor. Submitted to Climatic Change, Dec. 2005. Reviewer comments received and revisions underway; re-submit early 2007.

4. **Upland climate change impacts – towards improved site-scale assessments for land managers?**
John Coll, Stuart W. Gibb, John Harrison (2006). In *Global Change in Mountain Regions*. Ed. M.F. Price. Sapiens Publishing, Dumfries. pp 273-275.

Also provided are the conference posters which relate to the above outputs, these are;

Modelling climate change related shifts in land-use in the Scottish Highlands - an approach integrating local climatologies and climate model scenarios.
John Coll, Stuart W. Gibb and John Harrison. Presented at: Nature and People, conservation and management in the mountains of Northern Europe. Pitlochry, Scotland, UK. November 2002. Organised as part of the SNH contribution to the International Year of the Mountains.

North Atlantic Oscillation Driven Changes To Wave Climate in the Northeast Atlantic: Implications for Ferry Services to the Western Isles.
John Coll, David Woolf, Stuart Gibb, Peter Challenor and Michael Tsimplis. Presented at: 1. Human/Environment Interaction Research (HARC). 2. SEARCH (A Study of Environmental Arctic Change) Open Science Conference Seattle, Washington, October 2003. Attendance enabled by a Scholarship award from the Arctic Research Consortium of the United States (ARCUS).

Upland climate change impacts – towards improved site-scale assessments for land managers?
John Coll, Stuart W. Gibb, John Harrison. Presented at: Open Science Conference. Global Change in Mountain Regions: Perth, Scotland, UK, 1-5 October 2005. Extended abstract published in proceedings Spring 2006.