CHAPTER 2: **AN OVERVIEW OF CANADA'S CHANGING CLIMATE**

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1. INTRODUCTION

Warming of the global climate over the past century is unequivocal. It is evident in global atmospheric and oceanic temperature data, and from changes in a variety of other physical indicators, including declines in snow and ice cover. Emissions of greenhouse gases (GHGs) from human activity are the main cause of recent global warming and are expected to be the dominant cause of further warming over the coming century, the magnitude of which will be strongly influenced by whether anthropogenic emissions of GHGs continue to grow or are curtailed. These widely accepted conclusions are supported by a large body of evidence (e.g. IPCC, 2007; AMAP, 2011; NRC, 2011; IPCC, 2012, 2013) and provide the global context for examining current and projected changes in Canada's climate.

This chapter presents an overview of observed and future changes in a number of key indicators of the physical climate system (Figure 1) and provides context for the chapters that follow, highlighting significant advances in understanding since the 2008 Assessment Report ('From Impacts to Adaptation: Canada in a Changing Climate'; Lemmen et al., 2008). Focus is placed on the national scale, although regional trends and changes are described where suitable data exist. The Arctic receives particular attention due to the wealth of regional data that has been amassed for various cryospheric indicators. In addition, given the differing nature of the ocean basins bordering Canada, a regional perspective is necessary for discussion of changes in ocean climate, and in the section on freshwater resources, since available studies are primarily watershed-based.

It should be noted that this chapter does not provide climate change scenarios as a basis for the assessment in subsequent chapters. A few illustrative future climate change scenarios are presented for specific indicators and time periods. Also, this chapter makes only limited use of the results of the many new observational, modeling and analysis studies that have been appearing in the literature during the chapter's preparation, and which were synthesized in the Intergovernmental Panel on Climate Change (IPCC) Fifth Assessment Report, the first volume of which was released in 2013 (Stocker et al., 2013).

This chapter documents the evidence that shows that climate change in Canada is occurring, and discusses changes that are expected in the future, which will impact both natural and managed environments and the many economic and social activities that depend on them. This chapter provides a backdrop to what readers may experience in their own communities. It serves to illustrate the large forces at play, and how changes experienced locally will be nested within changes occurring within Canada, North America and globally.

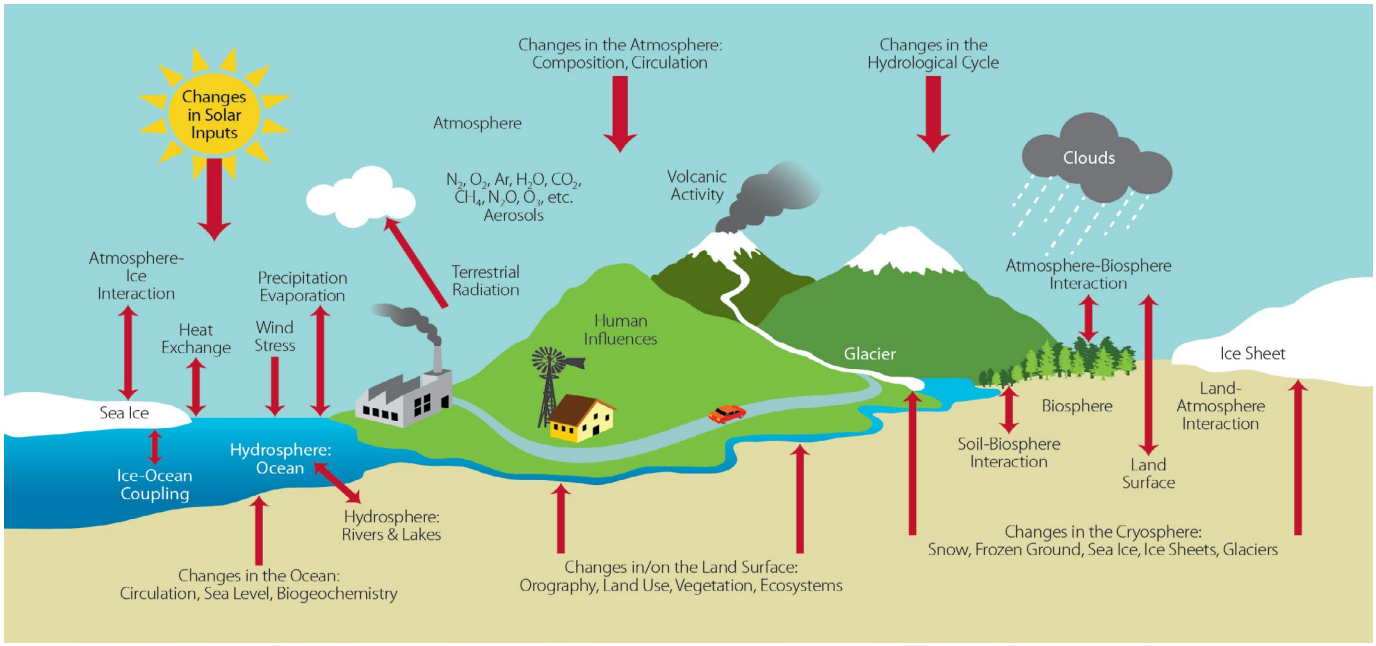

FIGURE 1: An illustration of the main components of the climate system (the atmosphere, hydrosphere [liquid water components], cryosphere [frozen water components], lithosphere [land surface] and biosphere [living things]) and the interactions between them. This chapter reports on four of the six changes shown in this schematic, excluding changes on the land surface and in solar outputs *(Source: Le Treut et al., 2007)*.

2. CHANGES IN AIR TEMPERATURE AND PRECIPITATION

2.1 OBSERVED CHANGES IN TEMPERATURE AND PRECIPITATION

2.1.1 TEMPERATURE

GLOBAL

A 100-year warming trend[1] of 0.74°C ± 0.18 °C in global surface air temperature was observed for the period 1906-2005 (IPCC, 2007). Several reports, including the World Meteorological Organization (WMO) Statement on the Status of the Global Climate (WMO, 2013) and the American Meteorological Society (AMS) State of the Climate Report (Blunden and Arndt 2012) have identified 2010, globally, as either the warmest or second warmest year on record (ranking varies very slightly with the methods used by different agencies to estimate global mean temperature) and have confirmed the existence of a strong, long-term global warming trend (Figure 2A , WMO, 2013). In addition, the decade 2001-2010 was the warmest on record, 0.21°C warmer than the previous decade, 1991-2000, which in turn was warmer than previous decades, consistent with a long-term warming trend (Figure 2B; WMO, 2011). Surface temperatures over land have warmed faster than over the oceans, with greatest warming over high northern latitudes (Trenberth et al., 2007). The Arctic continues to warm at about twice the rate of lower latitudes (Richter-Menge and Jeffries, 2011).

Natural fluctuations in climate can induce periods of a decade or two with little change in temperature, even with increasing levels of GHGs in the atmosphere (Easterling and Wehner, 2009). Despite the apparent slowdown in the rate of observed global warming over the last decade (*see* Figure 2A), twelve of the thirteen warmest years on record have been in the 21st century, the only exception being 1998 which was influenced by the strongest El Niño of the last century (WMO, 2013). The probability is very low that such a clustering of exceptionally warm years at the end of the observational record would occur in a climate that was not undergoing long-term warming (Zorita et al., 2008). When changes in natural factors known to influence short-term climate variability (e.g. volcanic eruptions and changes in solar radiation) are accounted for, the rate of global warming since 1980 has been shown to be steady (Foster and Rahmstorf, 2011).

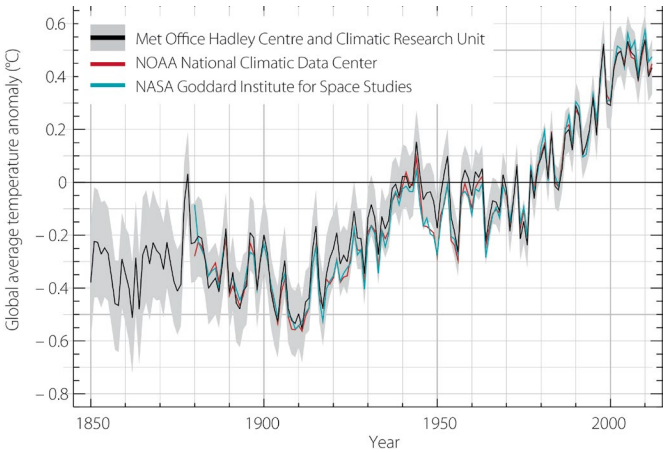

FIGURE 2A: Annual global average temperature anomalies (relative to 1961-1990) from 1850 to 2012 from the Hadley Centre/CRU (HadCRUT4) (black line and grey area, representing mean and 95% uncertainty range), the NOAA National Climatic Data Centre (red) and the NASA Goddard Institute for Space Studies (blue) *(Source: WMO, 2013).*

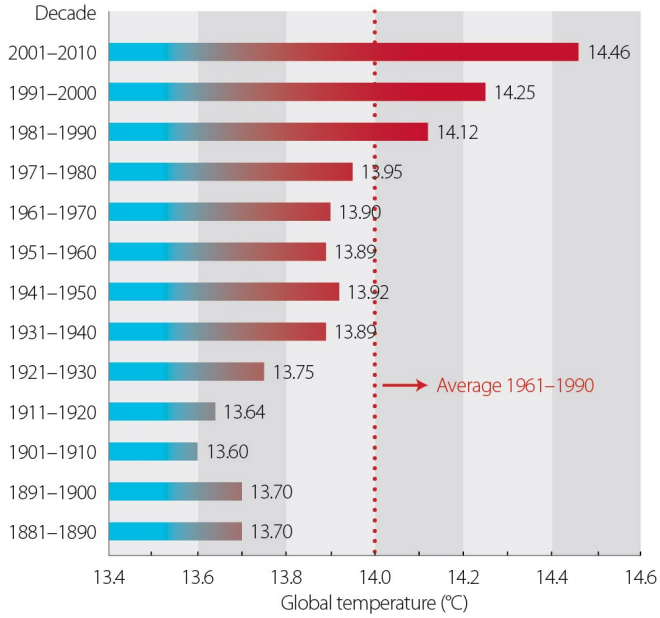

FIGURE 2B: Decadal global average combined land-ocean surface temperature (°C), combining three global datasets *(Source: WMO, 2011).*

[1] This long term trend was recently updated and is reported to be 0.85 [0.65 to 1.06°C] over the period 1880-2012 (IPCC, 2013).

Human activities affect the climate system by changing the land surface (e.g. deforestation) and altering the composition of the atmosphere. The latter involves increasing atmospheric concentrations of GHGs (that exert a warming influence) and also increasing concentrations of airborne particles (aerosols) that, for the most part, exert a climate cooling effect. Global warming since the mid-twentieth century has been attributed primarily to human emissions of GHGs (Hegerl et al., 2007). Similarly, a human influence on climate warming since the mid-twentieth century has been demonstrated for North America (Hegerl et al., 2007; Stott et al., 2010), Canada (Zhang et al., 2006) and over the entire 20th century for the Arctic (Gillet et al., 2008).

CANADA

The annual average surface air temperature over the Canadian landmass has warmed by 1.5°C over the period 1950-2010 (Figure 3; Vincent et al., 2012, *see* Box 1). Recent analysis shows that 2011 and 2012 were 1.5°C and 1.9°C warmer than the reference period (1961-1990 average); therefore, 2010 still stands as the warmest year on record in Canada, at 3.0°C above normal (Environment Canada, 2012).

While warming has been observed consistently across most of Canada, stronger trends are found in the north and west and warming has been weak along the Atlantic coast (Figure 4; *see also* Box 2). This regional pattern of stronger warming in the west vs. the east has been observed across North America and has been linked to shifts in large-scale atmosphere-ocean circulation patterns (Trenberth et al., 2007; *see* Box 3). Daily minimum temperatures in Canada have been rising slightly faster than daily maximum temperatures over 1950-2010. Warming in Canada is generally observed in all seasons (Vincent et al. 2012), with the greatest warming since 1950 occurring in winter and spring, with strong warming particularly evident in the western half of the country in these seasons (Figure 5). Warming trends are generally much weaker in the summer and fall; spatially, the summer warming is observed across the country, whereas during the fall, the warming is pronounced in the north and east (Figure 5). The spatial pattern of temperature changes during all seasons is consistent with previously reported patterns (Zhang et al., 2000), although the previously reported cooling in the northeast of the country is no longer evident in the longer time series due to recent warming in this region.

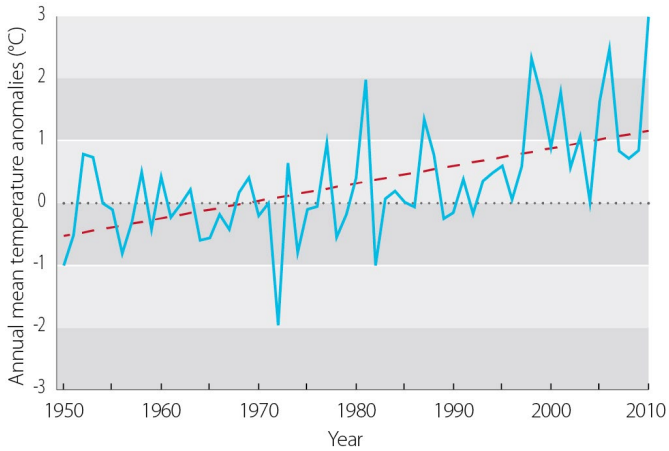

FIGURE 3: Annual mean temperature anomalies for Canada (°C), 1950-2010. Anomalies are calculated as departures from the 1961-1990 average (represented by zero on the Y-axis). A warming trend of 1.5°C over the period 1950-2010 is indicated by the red line *(Source: Vincent et al., 2012; Environment Canada, 2011).*

BOX 1

CREATING CLIMATE DATA SETS FOR LONG TERM TREND ANALYSIS

Canada's climate varies considerably from one region to another, and is characterized by significant variability – seasonally, from year to year, and over periods of multiple years. The challenge for climate change analysis is to detect a persistent trend (in annual or seasonal temperature or precipitation, for example) over space and time that may not be readily obvious from the 'noisy' data. To do this well, long term, continuous, data sets with good spatial coverage over the country are needed.

Since most climate stations in northern Canada were only established in the late 1940s, analysis of climate trends for the country as a whole is only possible from the second half of the 20th century onwards. Even over this period, there have been changes in observing practice, including instrumentation and instrument location, and some site locations have changed. These changes can generate artificial shifts in the data – referred to as "inhomogeneities" – that can interfere with climate trend analysis. In addition, there are known systematic biases in some instruments (e.g. wind-induced undercatch of precipitation by rain gauges) for which corrections are required. Reliable trend estimates can only be calculated when appropriate adjustments have been made to the original data to account for these methodological changes. This process is called "data homogenization". Stations selected for long term climate trend analysis are located outside major urban areas so no adjustments to the data are required to account for urban heat-island effects.

BOX 2

DEPICTING SPATIAL PATTERNS IN CLIMATE TRENDS ACROSS CANADA

Climate data are recorded at climate stations across the country and processed for long-term trend analysis (*see* Box 1). Spatial coverage of climate stations is uneven, with relatively few stations in northern Canada compared to southern Canada.

The maps of temperature and precipitation trends presented in this chapter (Figures 4, 5, and 7 to 10) provide information about the spatial coverage of the long-term stations, regional trends and the significance of those trends. Analysis and interpretation of these maps should focus on broad-scale patterns of change in and between different regions of the country, rather than on the trend values at specific sites. To illustrate future climate change, maps with continuous coverage (*see* Section 2.2) can be used because the climate models used to generate future climate simulations are grid-based and do not rely on input from observing networks. However, since variations at the scale of the computer grid and smaller are not resolved in these models, the focus of attention should still be on broad-scale patterns.

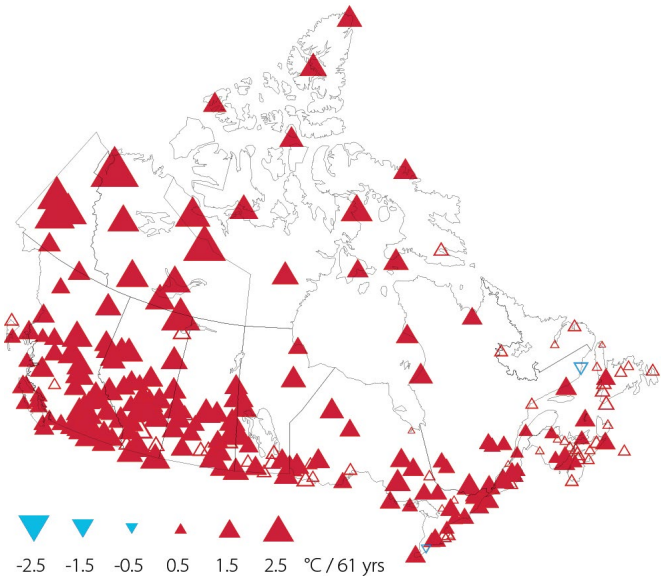

FIGURE 4: Trends in annual mean temperature for 1950-2010. Upward- (red) and downward- (blue) pointing triangles indicate positive and negative trends, respectively. Filled triangles correspond to trends significant at the 5% level. The size of the triangle is proportional to the magnitude of the trend. The legend may not include all sizes shown in the figure *(Source: Vincent et al., 2012).*

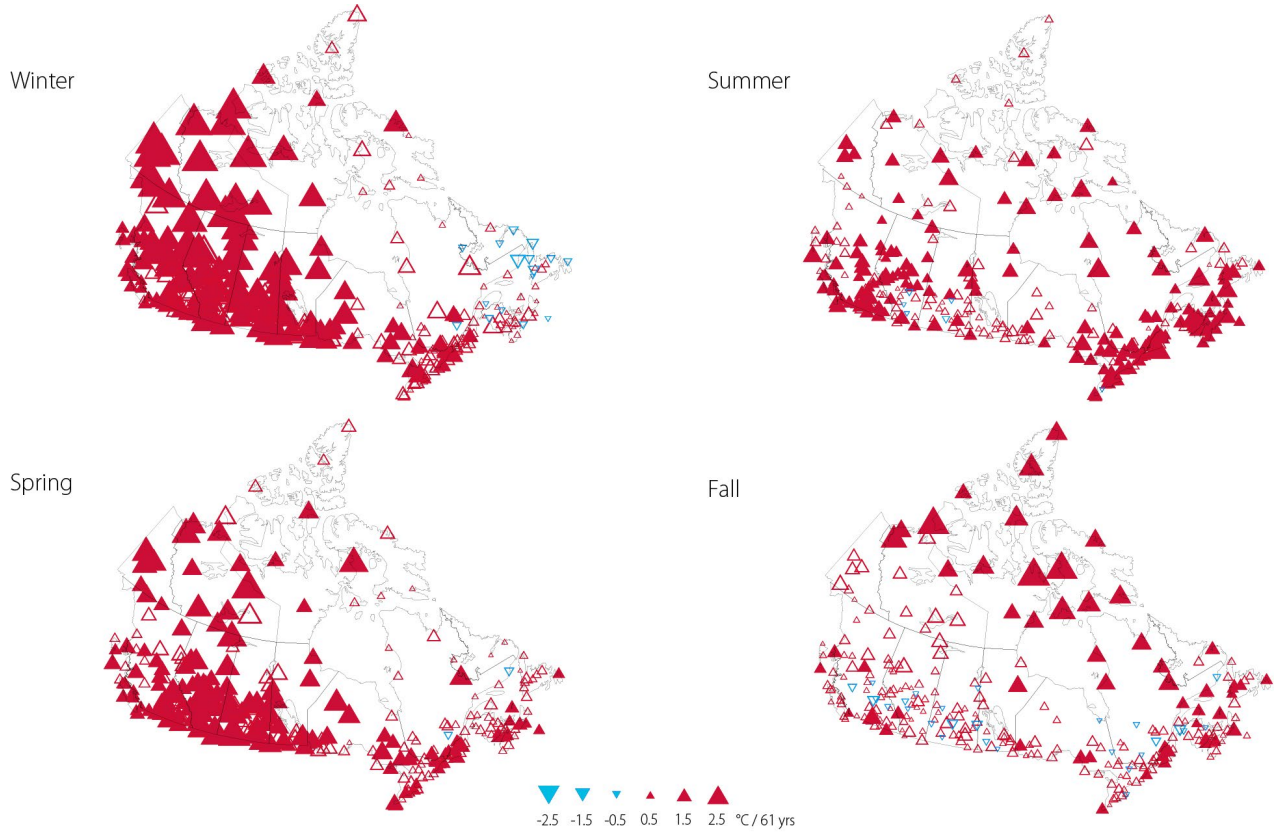

FIGURE 5: Trends in seasonal mean temperature for 1950-2010. Upward- (red) and downward- (blue) pointing triangles indicate positive and negative trends, respectively. Filled triangles correspond to trends significant at the 5% level. The size of the triangle is proportional to the magnitude of the trend. The legend may not include all sizes shown in the figure *(Source: Vincent et al., 2012.).*

BOX 3

INTERNAL CLIMATE VARIABILITY

The atmosphere, cryosphere, ocean and land are interconnected through the exchange of heat, freshwater, energy and gases, forming a coupled climate system. Because the cryosphere and ocean contain (or can absorb) large amounts of heat and freshwater, strong feedbacks can occur within this system, resulting in natural variations or "oscillations" sometimes referred to as internal climate variability. There is increasing evidence of the occurrence of these variations in modern observations, particularly on decadal time scales and space scales of ocean basins. Prominent examples include the Arctic and North Atlantic Oscillations (AO, NAO) in atmospheric pressure patterns, the Pacific Decadal and Atlantic Multidecadal Oscillations (PDO, AMO) in ocean surface temperature, and on shorter timescales, the El Niño/La Niña (warm/cool) variations in eastern tropical Pacific Ocean temperatures.

2.1.2 PRECIPITATION

Warming of the Earth's surface and atmosphere results in changes in evaporation and precipitation, and in atmospheric circulation patterns that influence where rain falls. In general, warmer temperatures lead to greater potential evaporation of surface water, thereby increasing the potential for surface drying and increasing the amount of moisture in the air. As warmer air can hold more moisture, more intense precipitation events are expected (Held and Soden, 2006; Trenberth, 2011). A shift in the latitudinal distribution of Northern Hemisphere precipitation has been observed, with increases at higher latitudes and decreases in the sub-tropics (Zhang et al., 2007; Min et al., 2008).

Precipitation trends are more difficult to detect than temperature trends (e.g. Trenberth et al., 2007; Warren and Egginton, 2008). Using adjusted daily precipitation data, Mekis and Vincent (2011a) show that Canada has generally become wetter in recent decades (an increase in annual precipitation of about 16% over the period 1950-2010; Figure 6). This increase is dominated by large changes in British Columbia and Atlantic Canada (Figure 7). For the past 61 years, 21% of the stations indicate statistically significant increases in annual total precipitation, while only a few significant negative trends are found throughout the country (Mekis and Vincent, 2011b). At most stations, total precipitation has increased in spring and fall, while many sites, especially those in western Canada, show declining winter precipitation (Figure 8). The observed decrease in total winter precipitation is mainly due to the decrease in winter snowfall, while winter rainfall has changed little (Mekis and Vincent, 2011a).

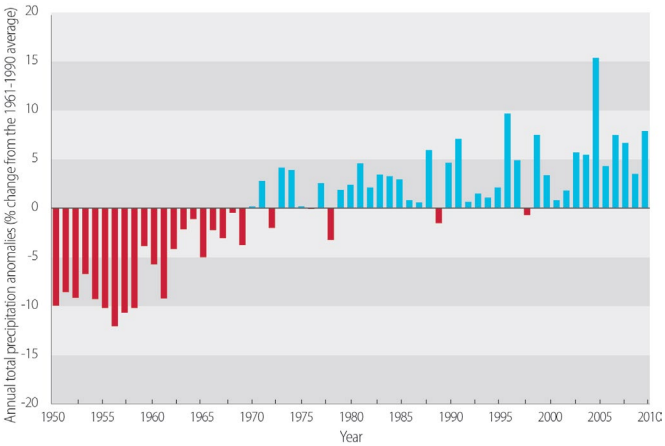

FIGURE 6: Annual total precipitation anomalies (expressed in % change from the 1961-1990 average) for Canada, 1950-2010 *(Source: Mekis and Vincent, 2011a; Environment Canada, 2011).*

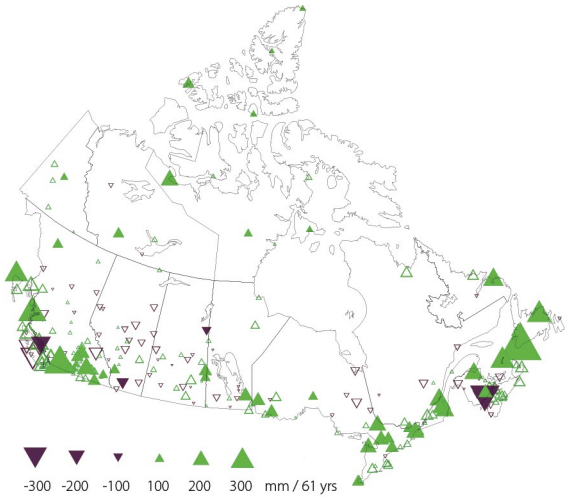

FIGURE 7: Annual total precipitation trends for 1950-2010. Upward- (green) and downward- (brown) pointing triangles indicate positive and negative trends, respectively. Filled triangles correspond to trends significant at the 5% level. The size of the triangle is proportional to the magnitude of the trend. The legend may not include all sizes shown in the figure *(Source: Mekis and Vincent, 2011b).*

Disaggregating total precipitation into rainfall and snowfall indicates that annual rainfall in Canada has increased by ~13% over 1950-2009 (Mekis and Vincent, 2011a). Trends indicate increasing rainfall across the country, although for many locations these trends are not statistically significant (Figure 9). Seasonally, many stations show significant increasing rainfall trends in the spring and fall.

Annual snowfall for Canada as a whole has increased by about 4% over 1950-2009 (Mekis and Vincent, 2011a), although many stations in western Canada show significant decreasing trends, while increasing trends occur in the north and Atlantic regions (Figure 9). Variability in winter precipitation, particularly in Western Canada, is strongly influenced by large-scale natural climate variations (such as the El Niño Southern Oscillation (ENSO) and the Pacific Decadal Oscillation (PDO) (*see* Box 3)), with below-normal precipitation associated with El Niño events and positive phases of the PDO, both of which have occurred more frequently since the mid-1970s (Bonsal and Shabbar, 2011). In several regions of southern Canada, there has been a shift in precipitation type, with decreasing snowfall and increasing rainfall (Figure 9), as would be expected with warming temperatures.

2.1.3 TEMPERATURE AND PRECIPITATION EXTREMES

In a changing climate, extreme temperatures and precipitation will also change as a result of shifts in mean conditions and/or as a result of changes in variability (Rummukainen, 2012). For example, warming is expected to be accompanied by a decrease in cold extremes and an increase in hot extremes. It is also expected that the global hydrological cycle will intensify with continued global warming, leading to an increasing intensity of both wet and dry extremes, and associated hazards such as floods and

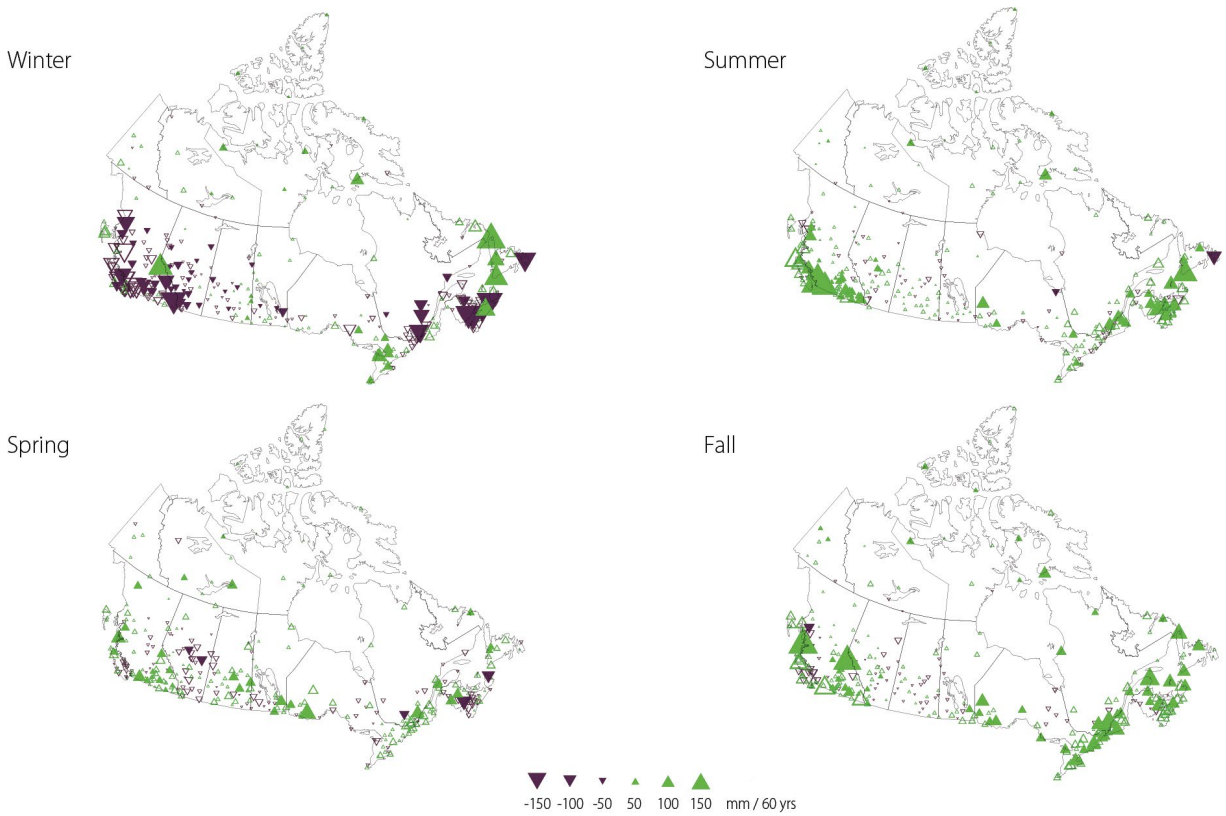

FIGURE 8: Seasonal total precipitation trends for 1950-2009. Upward- and downward-pointing triangles indicate positive and negative trends, respectively. Filled triangles correspond to trends significant at the 5% level. The size of the triangle is proportional to the magnitude of the trend. The legend may not include all sizes shown in the figure *(Source: Mekis and Vincent, 2011a)*.

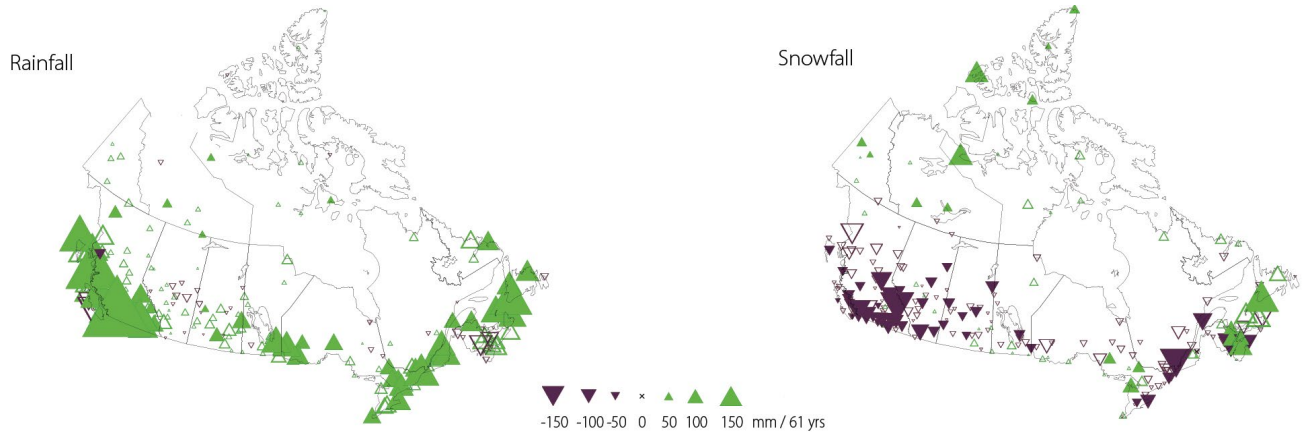

FIGURE 9: Annual rainfall and snowfall trends for 1950-2009. Upward- and downward-pointing triangles indicate positive and negative trends, respectively. Filled triangles correspond to trends significant at the 5% level. The size of the triangle is proportional to the magnitude of the trend. The legend may not include all sizes shown in the figure *(Source: Mekis and Vincent, 2011a)*.

droughts (Trenberth, 2011; Giorgi et al., 2011). An extreme event, by definition, is rare – making analysis of changes in extreme events challenging (Zhang et al., 2011a).

At the global scale, there is abundant evidence for changes in daily temperature extremes, with an increase in the number of warm days and warm nights and a decrease in the number of cold days and cold nights over much of the global land area (Seneviratne et al., 2012). An increase in the length and/or frequency of longer duration events (e.g. heat waves) was also found, although confidence in these results is lower than it is for changes in daily temperature extremes. There is generally less consistency and coherence in the patterns of observed changes in extreme precipitation relative to changes in extreme temperatures. Globally, however, there were more places showing an increase in the number of heavy precipitation events than a decrease, so the overall assessment was for an increase in heavy precipitation at the global scale (Seneviratne et al., 2012).

In Canada, temperature trends updated since the 2008 assessment indicate that cold events continue to decrease while warm events continue to increase. The frequency of cold nights during the winter (when the daily minimum temperature is below the daily 10th percentile) has decreased over 1950-2010 at most stations across the country; however, some small increasing trends are evident in southern Quebec and Atlantic Canada. Similarly, while the frequency of warm days during the summer (when the daily maximum temperature is above the daily 90th percentile) has increased nationally, small decreasing trends were found at several locations in the Canadian Prairies. At most stations, the annual frequency of cold nights has decreased and the annual frequency of warm days has increased, observations which are in agreement with an assessment of trends across North America (including Canada, the U.S. and Mexico; Peterson et al., 2008). Analysis to infer changes in the one-in-20 year extremes indicates that extreme minimum temperatures have warmed more than extreme high temperatures, and that the trends have been much stronger in the Canadian Arctic than in southern Canada (Wang et al., 2013). No updated Canada-wide evaluation of trends in summer heat waves or warm spells is available.

With respect to extreme precipitation in Canada, two indices, namely "very wet days" (number of days with precipitation ≥ 95th percentile value) and "heavy precipitation days" (number of days with ≥ 10 mm precipitation) have been recomputed to update the analysis presented in the 2008 Assessment to cover the period 1950-2010 (Vincent and Mekis 2006, updated; Figure 10). The results show patterns similar to earlier findings, with no consistent change in extreme precipitation for Canada as a whole. On a continental scale, while various indices of heavy precipitation have been increasing since 1950, the patterns have not been spatially uniform across

North America (Peterson et al., 2008). Increasing trends in precipitation intensity have been observed over about two-thirds of the northern hemisphere land area with sufficient data coverage for analysis (Min et al., 2011).

Drought is an extreme event with no standard definition, but in general refers to extended periods of abnormally dry weather that deplete water resources. A variety of indices are available for assessing changes in drought. Although drought occurs in most regions of Canada, much of the research has focused on the Canadian Prairies because the region is particularly susceptible to drought (Bonsal et al., 2011). Assessment of the variability of summer drought duration for the southern Prairies over different time scales indicates that 20th century droughts have been relatively benign compared to those that occurred in previous centuries (Bonsal et al., 2012). No trends in drought over the full 20th century have been discernible in any region of the country (Bonsal et al., 2011). Over the second half of the 20th century, regional trends towards more severe drought conditions were identified over southern and western Canada, as part of global analyses (Dai, 2011; Seneviratne et al., 2012).

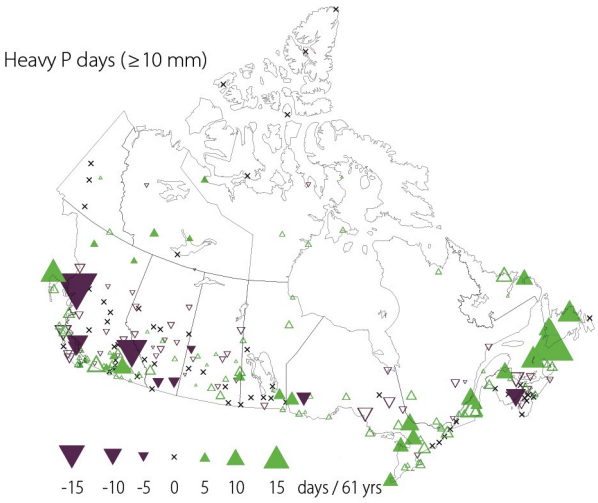

FIGURE 10: Trends in extreme precipitation for 1950-2010. Upward- and downward-pointing triangles indicate positive and negative trends, respectively. Filled triangles correspond to trends significant at the 5% level. The size of the triangle is proportional to the magnitude of the trend. The legend may not include all sizes shown in the figure. The symbol 'x' denotes a trend near zero *(Source: Vincent and Mekis, 2006, updated)*.

2.2 PROJECTED CHANGES IN TEMPERATURE AND PRECIPITATION

2.2.1 CLIMATE MODELS AND SCENARIOS

Climate change projections are provided by experiments run on supercomputers with mathematical models of the coupled atmosphere-ice-ocean-land system. These climate (or Earth) system models are based on physical laws governing the behavior of the system and interactions among its components, and can be used to simulate how the system will respond to forces of change. In such experiments, the models are driven by specified changes in 'climate forcers', including changes in GHG and aerosol emissions resulting from human activities. In general, researchers examine the results of experiments with many models using the same forcing scenario, as well as multiple forcing scenarios (*see* Box 4).

Based on SRES scenarios (Box 4), the IPCC Fourth Assessment Report provides central estimates and likely (66-100% probability) ranges of global average temperature change for the 2090-2099 period (relative to 1980-1999) of 1.8 °C (1.1 to 2.9°C) for B1, 2.8 °C (1.7 to 4.4 °C) for A1B and 3.4 °C (2.0 to 5.4 °C) for A2 (Meehl et al., 2007b). It further concluded that North America is very likely (90-100% probability) to warm this century and its annual mean warming is likely to exceed global mean warming in most areas, with strongest warming in winter and in northern regions (Christensen et al., 2007). The possibility of cooling in the northeastern part of Canada could not be ruled out because of the possible cooling in the North Atlantic associated with reduction in the Atlantic Meridional Overturning Circulation. Annual average precipitation is very likely to increase across Canada, while for southern Canada, precipitation is likely to increase in winter and spring but decrease in summer.

This chapter also presents selected climate change projections of surface air temperature change and precipitation for Canada based on the average results (computed means) from an ensemble of either 17 (B1) or 16 (A2) global climate models in CMIP3. These projections are intended to be broadly illustrative of the magnitude of potential changes across Canada over the course of this century in response to either a relatively low (B1) or relatively high (A2) forcing scenario. Recent global anthropogenic fossil fuel emissions more closely match the higher emission SRES scenarios (Peters et al., 2012). Readers are referred to Chapter 2 of the 2008 Assessment Report (Warren and Egginton, 2008) for a general discussion about climate modeling and impact assessment, and to the Canadian Climate Change Scenarios Network (cccsn.ec.gc.ca) for more technical information, including guidance on how to incorporate uncertainty in climate projections into adaptation planning.

BOX 4

CLIMATE CHANGE SCENARIO DEVELOPMENT

Future climate change scenarios presented in the 2008 assessment report (Lemmen et al., 2008) were based on a set of coordinated experiments undertaken by a small number of global climate modeling groups for the IPCC's Third Assessment Report (IPCC, 2001). The experiments were based on a subset of the emissions scenarios described in the IPCC Special Report on Emissions Scenarios (SRES) (Nakićenović et al., 2000 and summarized in Warren and Egginton, 2008). A new set of internationally coordinated climate change experiments was undertaken by a much larger number of climate modeling groups to support the IPCC's Fourth Assessment (IPCC, 2007). This set of coordinated experiments – the Coupled Model Intercomparison Project Phase 3 (CMIP3) experiments (Meehl et al., 2007a) – also used the SRES scenarios developed by the IPCC, with 3 scenarios routinely modeled by most groups, representing a range of future anthropogenic radiative forcing from low (B1) to medium (A1B) to medium-high (A2). While these three scenarios capture a substantial portion of the range of projected emissions from the larger set of SRES scenarios, they do not capture the full range. Furthermore, none of the SRES scenarios included explicit consideration of climate change mitigation efforts, although they did capture future worlds with relatively low to relatively high emissions even in the absence of specific policies to address emissions.

The most recent coordinated global climate Coupled Model Intercomparison Project (CMIP5) used a new set of scenarios as the basis for projecting future climate change (Moss et al., 2010; Hibbard et al., 2011; van Vuuren et al., 2011; Taylor et al., 2012). The new scenarios, referred to as Representative Concentration Pathways (RCPs), describe trajectories of atmospheric concentration over time (for GHGs, aerosols and other air pollutants, and the resulting trajectories in net radiative forcing). Four scenarios of interest to the policy and scientific communities were chosen to represent a range of radiative forcing over the 21st century, encompassing scenarios assuming ambitious mitigation to those assuming little. Hence, RCPs differ from SRES scenarios by explicitly considering greenhouse gas mitigation efforts.

The CMIP5 experiments, based on the new RCPs, have provided a focus for discussion of future global, continental and regional-scale climate change in the Fifth Assessment Report of IPCC Working Group I. Simulations from CMIP5 were not available during the original drafting of this chapter, nor has the impacts and adaptation community had much time to conduct research using these new projections. New climate change scenarios for Canada based on these experiments will be made available to the Canadian research community and to interested Canadians through the Canadian Climate Change Scenarios Network at cccsn.ec.gc.ca.

2.2.2 CHANGES IN SEASONAL TEMPERATURE AND PRECIPITATION

Scenario maps of projected seasonal changes in surface air temperature (Figure 11) and precipitation (Figure 12) are presented for the middle and end of the century, relative to 1961-1990 averages, based on the CMIP3 multi-model mean results for the low (B1) and medium-high (A2) SRES emissions scenarios (*see* Box 4). These projections are generally consistent with those presented in the 2008 Assessment Report.

Seasonal variation in warming patterns are evident, with largest increases in winter occurring at high latitudes (northern Canada), while in summer the greatest warming occurs at mid-latitudes (southern Canada). Summertime warming is generally projected to be more uniform across the country, with the largest changes projected for the continental interior. Stronger latitudinal gradients are apparent in autumn, and especially so in winter. Ensemble mean results over the whole domain show that even under the low emission scenario (B1), by the middle of the century all of Canada is projected to warm by about 1.5 to 2.5°C in the season of weakest projected warming (summer). Average wintertime temperatures across most of Canada are projected to increase by ~3 to 7°C under the A2 scenario towards the end of the century, with warming of > 9°C projected in and around Hudson Bay and the High Arctic (Figure 11, panel P). This pattern of enhanced warming at high northern latitudes is a near-universal feature of climate model projections under a range of emissions scenarios, and is strongly linked to reduced snow and sea ice cover (Serreze and Barry, 2011).

Projections of precipitation change are generally less robust than those for temperature, exhibiting greater variability among models. Increases in precipitation are projected for the majority of the country and for all seasons, the exception being parts of southern Canada where a decline in precipitation in summer and fall is projected (Figure 12, panels F-L). Even in areas where summer precipitation increases, higher evaporation rates associated with warmer summers will increase the tendency towards drier conditions. An increase in aridity in southern Canada is projected, with large variations between scenarios (Sheffield and Wood, 2008; Dai, 2011).

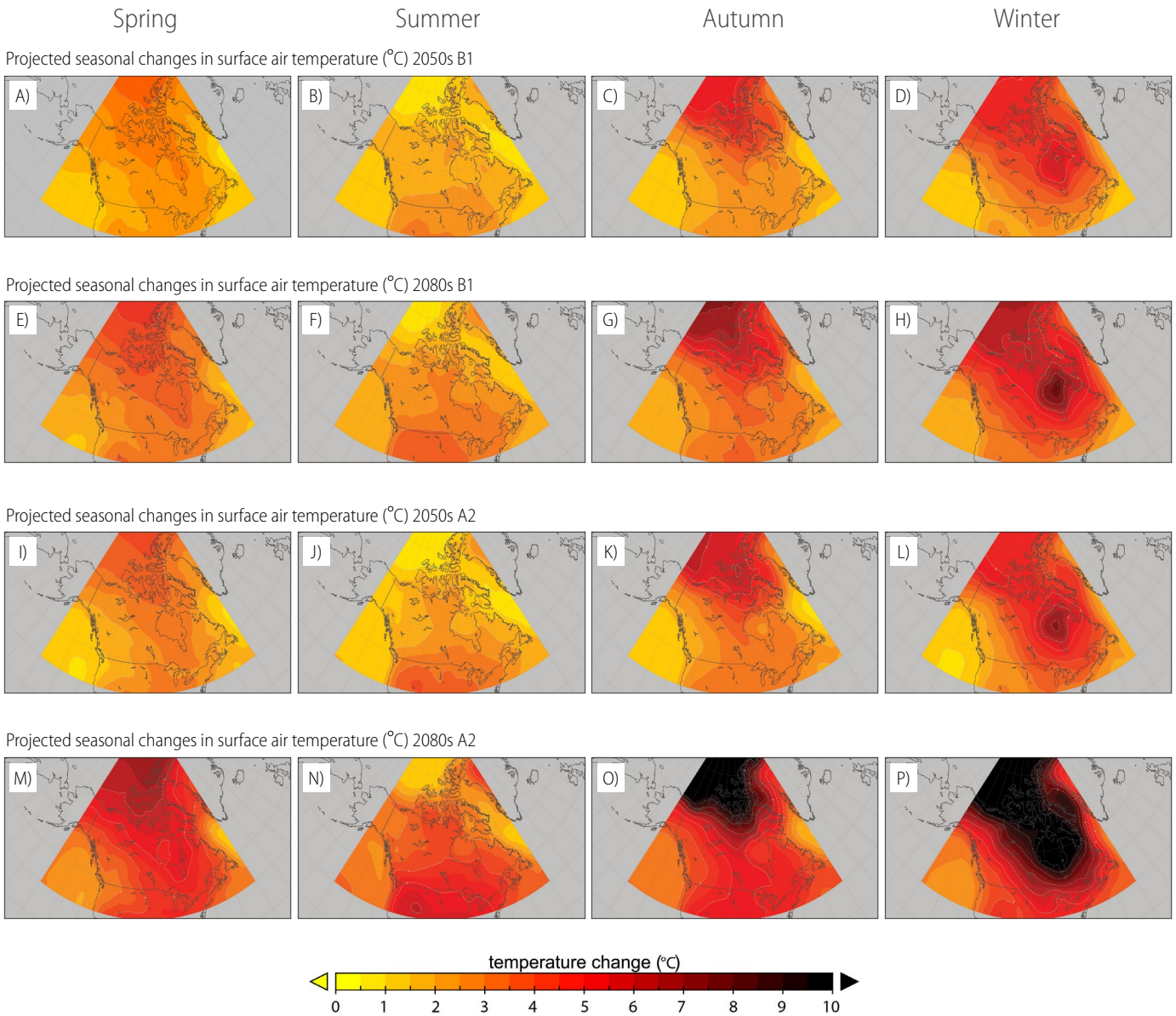

FIGURE 11: Projected seasonal changes in temperature across Canada for the middle and end of the 21st century under various SRES scenarios. Changes are expressed relative to average values between 1961-1990. Row 1 (A-D) is scenario B1 mid-century, row 2 (E-H) is B1 towards the end of the century, row 3 (I-L) is A2 mid-century, and row 4 (M-P) is A2 towards the end of the century. Column 1 (A, E, I, M) is Spring, Column 2 (B, F, J, N) is Summer, Column 3 (C, G, K, O) is Autumn, Column 4 (D,H, L, P) is Winter *(Source: Canadian Centre for Climate Modeling and Analysis)*.

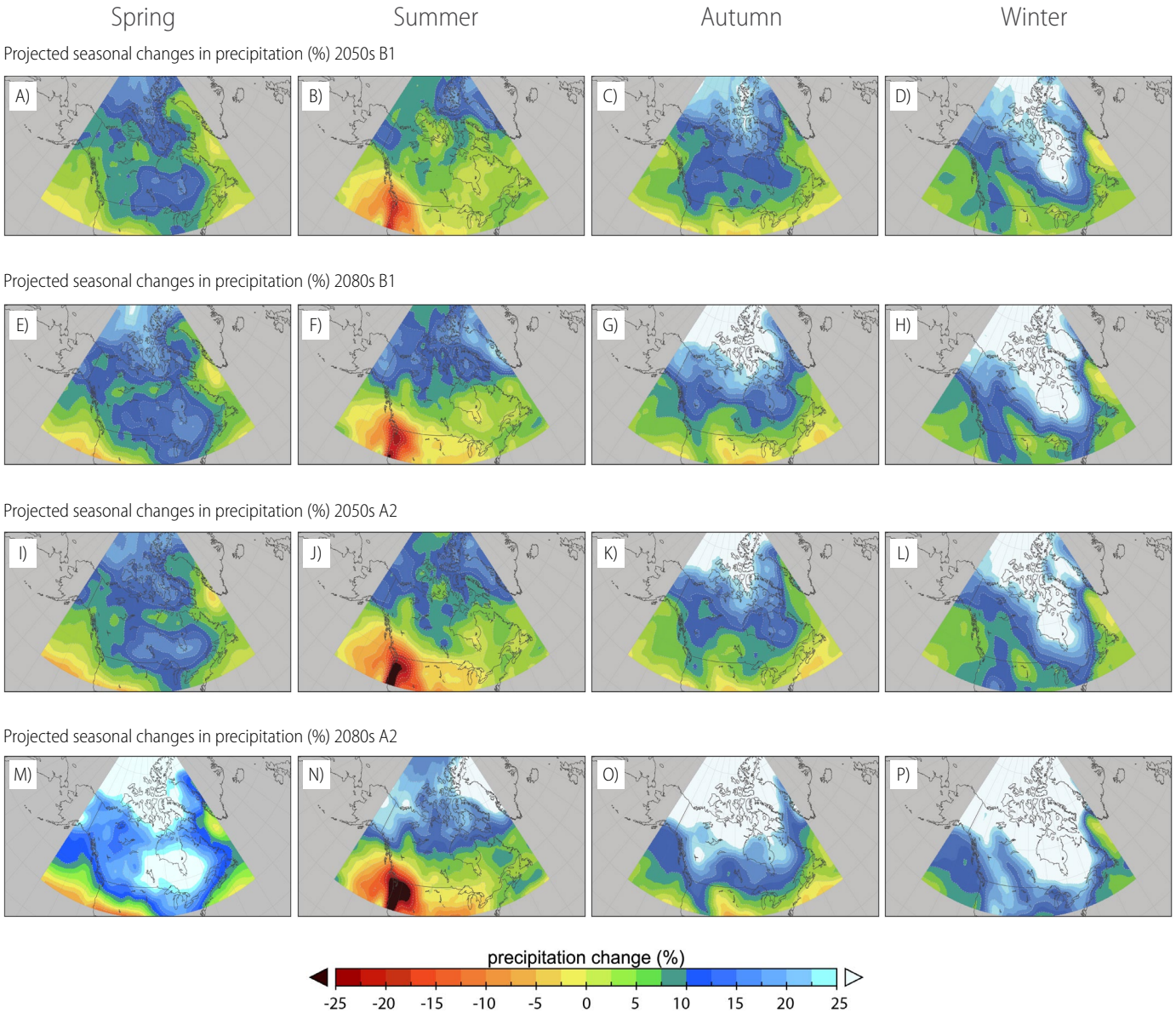

FIGURE 12: Projected seasonal changes in precipitation across Canada for the middle and end of the 21st century under various SRES scenarios. Changes are expressed relative to average values between 1961-1990. Row 1 (A-D) is scenario B1 mid-century, row 2 (E-H) is B1 towards the end of the century, row 3 (I-L) is A2 mid-century, and row 4 (M-P) is A2 towards the end of the century. Column 1 (A, E, I, M) is Spring, Column 2 (B, F, J, N) is Summer, Column 3 (C, G, K, O) is Autumn, Column 4 (D,H, L, P) is Winter *(Source: Canadian Centre for Climate Modeling and Analysis)*.

2.2.3 CHANGES IN TEMPERATURE AND PRECIPITATION EXTREMES

Changes in extreme events are of particular concern for adaptation planning. The IPCC Special Report on Managing the Risks of Extreme Events and Disasters to Advance Climate Change Adaptation (IPCC, SREX; 2012) provided a comprehensive assessment of projected changes in extreme weather and climate events, such as floods and droughts – at global, continental, and regional scales – using literature that for the most part was based on CMIP3 modeling experiments (Box 4) (Seneviratne et al., 2012). The report concludes that it is virtually certain (99 to 100% probability) that increases in the frequency and magnitude of warm days and nights, coupled with decreases in the frequency and magnitude for cold days and nights will occur globally during the 21st century and that the length, frequency and/or intensity of warm spells, including heat waves, are very likely to increase over most land areas (Seneviratne et al., 2012). Kharin et al. (2007) concluded that the return period for what is currently a one-in-20-year[2] extreme hot day would become a one-in-5 year event over most of Canada by mid-century. By the end of the century, such warm events are projected to become even more commonplace (Seneviratne et al., 2012; Gutowski et al., 2008).

With respect to precipitation, Kharin et al. (2007) found that return periods for one-in-20-year extreme daily precipitation events would become a one-in-10 year event by mid-century for mid to high latitude regions under moderate to high emission scenarios. The SREX concluded that the frequency of heavy precipitation is likely to increase in the 21st century over many areas of the world, but also emphasized the relatively large uncertainties associated with projections of extreme precipitation (Seneviratne et al., 2012). Regional climate models or statistical downscaling approaches can reveal important details about spatial patterns of change not evident in global studies. For example, the Canadian Regional Climate Model (CRCM) was used to explore future changes in warm-season (April-September) single and multi-day extreme precipitation events using the A2 emission scenario (Mladjic et al., 2011). In addition to finding an increase in future 20-year (and longer) return values of 1 to 7 day precipitation extremes for most parts of Canada (i.e. the 20-year events were projected to have larger precipitation totals), the study also concluded that the CRCM underestimates precipitation extremes over most of Canada, when evaluated against observed changes.

In assessing the risk of changes in drought, available studies suggest a strong tendency towards reduced aridity in winter and increased aridity in summer over large areas of the Canadian landmass. However, the lack of model agreement in the direction of projected change over many areas of Canada, including south-central Canada, indicates that these results should be interpreted cautiously (*see* Figure 3.9 in Seneviratne et al., 2012). Choice of drought index has also been shown to influence the results (Bonsal et al., 2012). The CRCM was used to evaluate potential changes in the frequency of dry days across Canada for the April–September period under the SRES A2 emission scenario, with increases in the mean number of dry days and in maximum dry spell duration (for given return periods) demonstrated for parts of southern Canada. The southern Prairies were identified in particular as having a higher likelihood of drought conditions in the future (Sushama et al., 2010, *see also* Bonsal et al., 2012).

3. CHANGES TO THE CRYOSPHERE

Canada is a northern country, and snow and ice are the dominant land cover for much of the country for much of the year. The presence or absence of snow and ice on and below the surface, and its seasonal variation, play important roles in climate from local to global scales. Changes to components of the cryosphere – sea ice, freshwater (lake and river) ice, snow cover, glaciers, ice caps, ice sheets and permafrost – are important indicators of a changing climate due to their climatic sensitivity and the importance of related impacts. For example, the cryosphere includes important stores of fresh water in the form of glaciers, ice caps and ice sheets. When the melt rate exceeds the accumulation rate, which is currently the case over most of Canada, the water released contributes to sea level rise. When ice is within frozen soils (permafrost), melting can lead to collapse of the soil structure, with consequences for overlying infrastructure and local hydrology. Water supply and soil moisture in many parts of the country can be affected by changes in both snow and glacier cover, with runoff from spring melt often critical to surface water supply in the high demand summer period. Loss of Arctic sea ice will have direct consequences for residents of the Arctic, for whom many aspects of community life and economic activity will be severely affected. As well, decreased sea ice will increase access to the Arctic Ocean and trans-Arctic shipping, with social, economic and environmental consequences.

[2] A one-in-20 year event means that such an event has, on average, a 5% chance of occurring in any given year. Similarly, a one-in 2-year event has, on average, a 50% chance of occurring in a given year.

Arctic research in Canada and other countries was substantially expanded during the third International Polar Year (IPY; 2007-2008). This work, and other recent reports on Arctic environmental change provide updated information on trends in various cryospheric indicators (AMAP, 2011; Arctic Report Card, 2012; Derksen et al., 2012). This is part of a growing body of evidence documenting widespread reductions in the spatial extent and mass of the cryosphere in response to warming air temperatures across the circumpolar north. Updated information on a few key indicators, with a focus on results for Canada, is provided here.

3.1 PERMAFROST

Permafrost underlies the northern half of the Canadian landmass, with a zone of relatively thin, warm, discontinuous permafrost lying south of the larger, continuous permafrost zone that extends up to the high Arctic (Smith, 2011). In the continuous permafrost zone, permafrost can extend tens to hundreds of metres below the surface. Ground temperature below the depth of seasonal variation is a good indicator of decade-to-century-scale climate variability (Romanovsky et al., 2010). Permafrost temperatures measured in boreholes at numerous sites across Canada have all increased over the past two to three decades (Smith et al., 2010; Figure 13). The magnitude of warming varies from one region to the next, reflecting differences in climate as well other factors including elevation, snow cover and the physical properties of the permafrost itself. In general, the temperature of colder permafrost has increased more rapidly than that for warmer permafrost. This difference is linked in part to the lack of vegetation and thick snow cover at high latitudes, which, south of the treeline serve to insulate the ground from air temperatures (Romanovsky et al., 2010). For permafrost at temperatures close to 0°C, energy is also being utilized to convert ice to water (phase change) rather than for temperature change.

Across northwestern Canada, warming of permafrost temperatures has been fairly continuous over the past 20 to 30 years, with evidence that warming rates have recently slowed. Across the eastern Arctic and northern Quebec, warming did not begin until 1993 and has occurred quite rapidly since then (Smith et al., 2010; *updated in* Derksen et al., 2012). Permafrost temperatures have risen about 0.2°C/decade on average in warm, discontinuous permafrost regions. There are fewer measurement sites farther north in cold tundra regions, but increases of ≥1°C/decade in permafrost temperatures have been recorded since the mid-1990s (Smith et al., 2010; *updated in* Derksen et al., 2012). These temperature trends for two representative sites, Norman Wells (warm permafrost) and CFS Alert (cold permafrost), are illustrated in Figure 13. The observed increases in permafrost temperature are largely attributable to increases in winter air temperature (Smith et al., 2012).

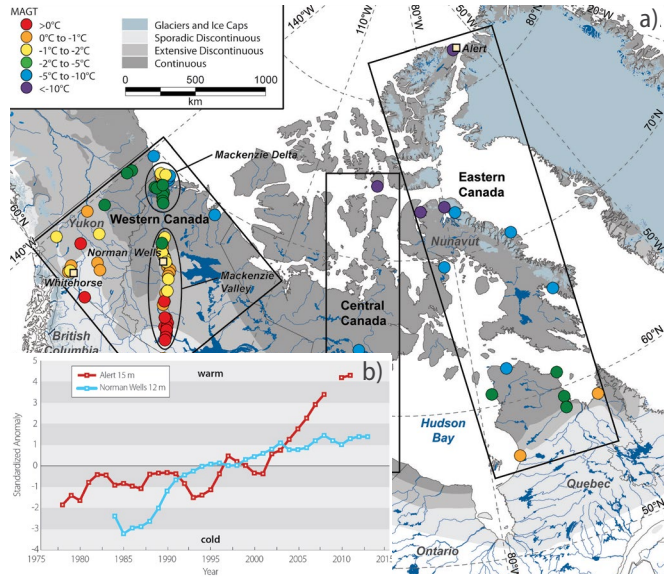

FIGURE 13: a) Mean annual ground temperature (MAGT) recorded during the IPY (2007-09) (from Smith et al., 2010). MAGT is determined at the depth of zero annual amplitude (depth to which seasonal variation penetrates) or the nearest measurement depth to it. Permafrost zones are from Heginbottom et al. (1995). **b)** Permafrost temperature standardized anomaly time series *(from Derksen et al., 2012)* relative to 1988-2007 mean for a site near Norman Wells (depth 12 m) in the central Mackenzie Valley and CFS Alert Nunavut (depth 15 m) in the high Arctic *(Credit: Sharon Smith, Natural Resources Canada)*.

With Arctic warming amplified compared to the global average, it is expected that warming of the permafrost will continue and that permafrost temperatures may increase more rapidly in the future than has been observed in records to date. However, even at the rapid warming rates seen in colder permafrost, the low average temperatures of much of the permafrost in the Arctic mean that it will take many decades to centuries for colder permafrost to completely thaw (Smith et al., 2010). Therefore, while warm, thin permafrost could eventually disappear, in colder permafrost regions climate warming will lead to a thickening of the active layer (seasonally thawed surface layer) and a decrease in permafrost thickness (Callaghan et al., 2011a).

The consequences of thawing permafrost are expected to be widespread for both natural ecosystems and human communities. Among these concerns are those related to the changing hydrology of northern ecosystems, decomposition of previously frozen soil carbon and the associated release of GHGs, and the loss of structural support from frozen ground with impacts on transportation and other infrastructure (Callaghan et al., 2011a; *see also* Chapters 3 and 8 of this report).

3.2 SNOW COVER

Year-to-year variability in snow cover, measured as snow cover extent (SCE) and snow cover duration (SCD), is closely linked to air temperature, particularly during the spring melt season when surface albedo feedbacks are strongest (Brown et al., 2010; Brown and Robinson, 2011). Changes in satellite-derived spring Arctic snow cover over the period 1967-2012 show statistically significant reductions over both North America and Eurasia in May and June (when snow cover is mainly located over the Arctic) (Derksen and Brown, 2011; Derksen and Brown, 2012). Successive records for the lowest June snow cover extent have been set each year for Eurasia since 2008, and in 3 of the past 5 years (2008-2012) for North America. The loss of June snow cover extent between 1979 and 2012 (-18% per decade relative to the 1979-2000 mean) is greater than the loss of September sea ice extent (-13% per decade) over the same period (*see* Section 3.5). Observed reductions in June SCE over the past decade now exceed the minimum June SCE simulated by an ensemble of climate models for this time period (Derksen and Brown, 2012).

Statistically significant negative trends have also been observed during spring over the Canadian land mass (Statistics Canada, 2012; Figure 14) with declines in snow cover of 7%, 13% and 34% in April, May and June, respectively, over the 1972-2010 period. Regionally, significant decreases in spring snow cover due to earlier melt have been observed over western and northern Canada (Figure 15, Zhang et al., 2011b), consistent with seasonal warming trends in these regions. Trends in spring snow cover extent for Canada are part of a larger-scale northern hemispheric trend of declining spring snow cover, which has become more rapid in recent decades, relative to rates of decline seen in long-term data sets covering the period 1922-2010 (Brown and Robinson, 2011).

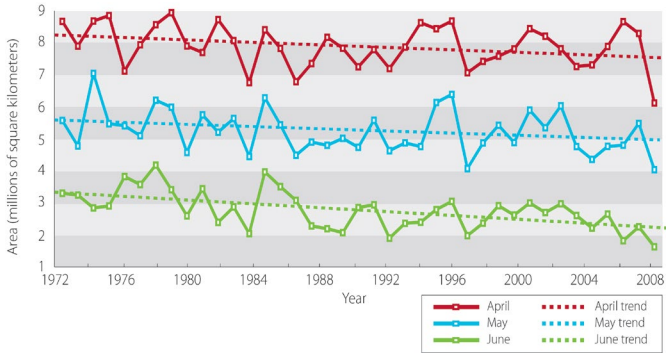

FIGURE 14: Changes in spring season snow cover extent over the Canadian land mass, 1972-2010, for the months of April (red lines), May (blue lines) and June (green lines) *(Source: Statistics Canada, 2012).*

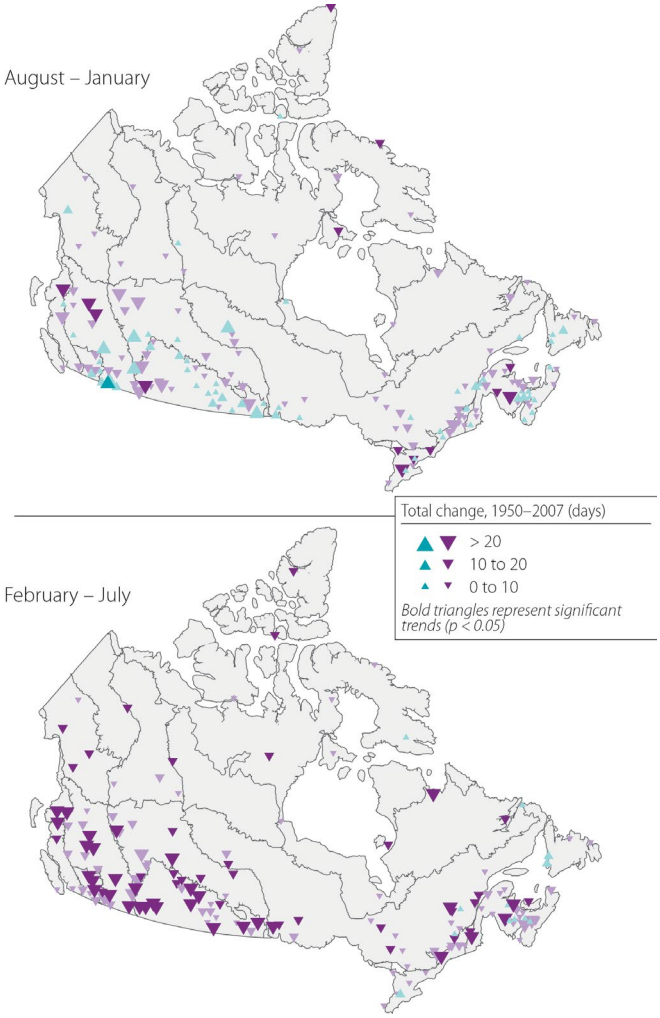

FIGURE 15: Change in the number of days with ≥ 2 cm of snow on the ground, 1950-2007, in a) the first half of the snow season (August to January) which indicates changes in the start date of snow cover, and b) in the second half of the snow season (February to July) which indicates changes in the end date of snow cover. Data are from daily snow depth observations. Boundaries on the map represent terrestrial ecozones *(Source: Zhang et al., 2011b)*.

In Canada, no country-wide trend in fall snow cover duration is evident in daily snow-depth observations (Figure 15), although some stations report locally significant trends toward later snow cover onset in the fall. There is, however, evidence of significant trends toward later freeze-up and snow cover onset over the Arctic since 1979 (e.g. Markus et al., 2009; Liston and Hiemstra, 2011).

Challenges in developing scenarios of future changes in snow cover using global climate models relate both to limitations of current models to capture processes important to the evolution of the snow pack, as well as limitations in

observations needed for comprehensive evaluation of the models (Callaghan et al., 2011b). Widespread decreases in the duration of snow cover are projected across the Northern Hemisphere (Figure 16; Raisanen, 2008; Brown and Mote, 2009) with the largest changes in maritime mountain regions, such as the west coast of North America. This is related to the high sensitivity of winter precipitation in these regions to small changes in temperature (Brown and Mote, 2009). Climate models also project increases in maximum snow accumulation over northern high latitudes in response to projected increases in cold season precipitation.

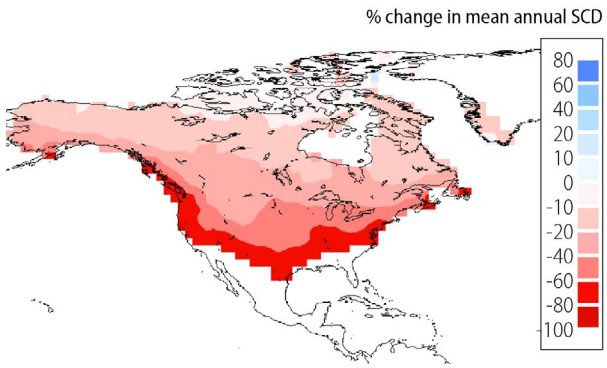

FIGURE 16: Projected percent changes in mean annual snow cover duration (SCD) between 1970-1999 and 2070-2099 for North America, using the IPCC SRES A2 emission scenario, for the eight global climate models used by Brown and Mote (2009, Fig. 10) *(Source: Ross Brown, Climate Research Division).*

3.3 GLACIERS

Many glaciers can be considered remnants from the last Ice Age that ended ca. 10 000 years ago, and as such have been in slow (and not necessarily continuous) retreat from their maximum positions for millennia. However, glaciers can also undergo quite rapid changes in response to changes in temperature and precipitation. The accelerated shrinkage of glaciers evident in many locations during the late 20th century has likely occurred in response to warming over this time period (Lemke et al., 2007). Glacier mass balance (the difference between the gain in mass from snow and ice accumulation and the loss of mass from melting and iceberg calving) is a sensitive indicator of climate change and tracks changes in glaciers over time scales of years to decades.

Glaciers in Canada are found primarily in two locations: the mountains of the western Cordillera, stretching from the Yukon to southern British Columbia and Alberta; and along the eastern margin of the Canadian Arctic Archipelago in the High Arctic. Glaciers from both regions show statistically significant declines in mass over the decades since ~1960 (Figure 17; WGMS, 2011). Multi-decadal time series show that changes have been larger for glaciers in the Western Cordillera region (Peyto, Place and Helm sites) than for those in the High Arctic (Devon NW, Meighton and White sites) (Figure 17). Cordilleran glaciers are generally exhibiting strongly negative mass balances (net losses in mass) and shrinking rapidly to the smallest extents in several millennia (Demuth et al., 2008). For example, assessment of multi-decadal changes in Yukon glaciers estimated that 22% of the surface area of these glaciers has been lost over the 50-year period since 1957-58 (Barrand and Sharp, 2010). Glaciers in British Columbia and Alberta have lost about 11% and 25% of their surface areas respectively, over the period 1985-2005 (Bolch et al., 2010).

In the High Arctic, glaciers are thicker and larger in area and thus respond more slowly to regional changes in climate. While these glaciers have exhibited more modest losses of mass compared to Cordilleran glaciers, significant negative mass balances are evident from the early 1960s into the first decade of the 21st century (Figure 17; Statistics Canada, 2010; WGMS, 2011). The rate of mass loss for glaciers throughout the High Arctic has increased sharply since 2005, in direct response to warm regional summer temperatures (Gardner et al., 2011; Sharp et al., 2011a). The mean rate of mass loss from four monitored glaciers in the Queen Elizabeth Islands between 2005-2009 was nearly 5 times greater than the ~40 year average over the period 1963-2004 (Sharp et al., 2011a). The recent series of warm summers in Arctic Canada has also been associated with major break-up events for the ice shelves that fringe northern Ellesmere Island (Copland et al., 2007; Sharp et al., 2011b; Pope et al., 2012).

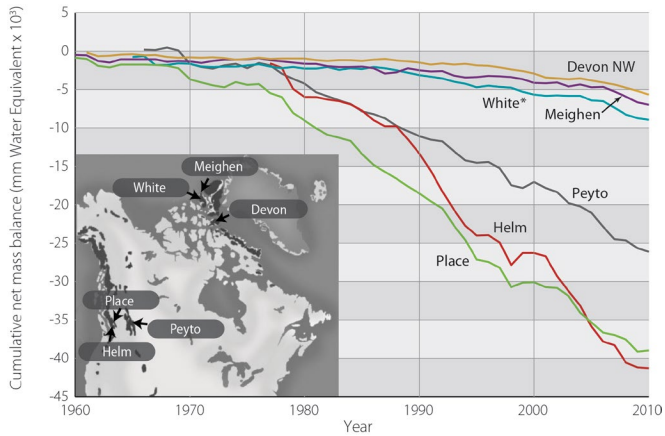

FIGURE 17: Cumulative net glacier mass balance patterns since observations first began over reference observing sites in the Canadian High Arctic (Devon and Meighen Ice Caps, and White Glacier*), the Canadian Rocky Mountains (Peyto Glacier) and the Cascade and Southern Coast Mountains (Place and Helm Glaciers). Values represent thickness change averaged over the glacier area (in mm water equivalent x 10^3). Inset: The location of each site in Panel A. *White Glacier data courtesy of Trent University *(Source: Natural Resources Canada, Geological Survey of Canada).*

Changes in the mass balance of glaciers in the Canadian High Arctic are driven mainly by changes in summer melt, as there is little interannual variability in snow accumulation. The high temperature-sensitivity of High Arctic glaciers will more than counteract projected precipitation increases, such that continued wastage of glaciers with ongoing warming is expected. Arctic glaciers are projected to be significant contributors of land ice to 21st century sea level rise (Gardner et al., 2011; Radić and Hock, 2011; Lenaerts et al., 2013; *see also* Section 5.4). Ongoing reductions in glacier extent are expected to reduce run-off in the longer term and impact water availability in glacier-fed rivers with consequences for aquatic habitat and for various human activities, including hydro-electricity generation (Sharp et al., 2011a).

3.4 FRESHWATER ICE

Using the historical record of ice cover from 1950-2005, trends towards earlier ice-free dates have been observed for most of the country but are particularly evident in Western Canada (Figure 18) (Duguay et al., 2006; Latifovic and Pouliot, 2007). Lake ice freeze-over dates have shown few significant trends (Duguay et al., 2006). Trends towards earlier break-up in river ice have also been observed through the latter half of the 20th century, again particularly in Western Canada (Beltaos and Prowse, 2009). River ice freeze-over was found to be spatially complex, and while some locations have shown later freeze, there have been no consistent trends throughout Canada (Beltaos and Prowse, 2009).

Large-scale internal climate variability (*see* Box 3) also influences changes in Canada's lake ice cover (e.g. Bonsal et al., 2006; Wang et al., 2012). The strongest effects are typically due to Pacific variability, although the extreme eastern parts of Canada are more affected by the North Atlantic Oscillation, and the Great Lakes region is influenced by both Pacific and Atlantic variability (Brown and Duguay, 2010).

Lake ice models driven by the CRCM (A2 scenario, *see* Box 4) provide projections of future ice cover conditions for the mid-21st century throughout Canada. Generally, ice cover break-up dates are expected to advance in the range of 1 to 3-1/2 weeks, while freeze-up dates are expected to be delayed by up to 2 weeks. The resulting ice cover duration is expected to decrease by up to a month depending on the depth of the lake, with greater reductions found for deeper lakes (Brown and Duguay, 2011; Dibike et al., 2012). River ice duration is also predicted to decrease by approximately three weeks, based on predicted changes to the 0°C isotherm (Prowse et al., 2007). Maximum lake ice thickness is predicted to decrease in the range of 10 to 30 cm, with loss exceeding 40 cm in some Arctic areas (Brown and Duguay, 2011; Dibike et al., 2012). Changes in ice thickness are also affected by snow cover, with ice thickness showing greater reductions when no snow cover is present. A severe reduction in the number of areas where occasional summer or perennial ice cover is found throughout the High Arctic is also projected (Brown and Duguay, 2011).

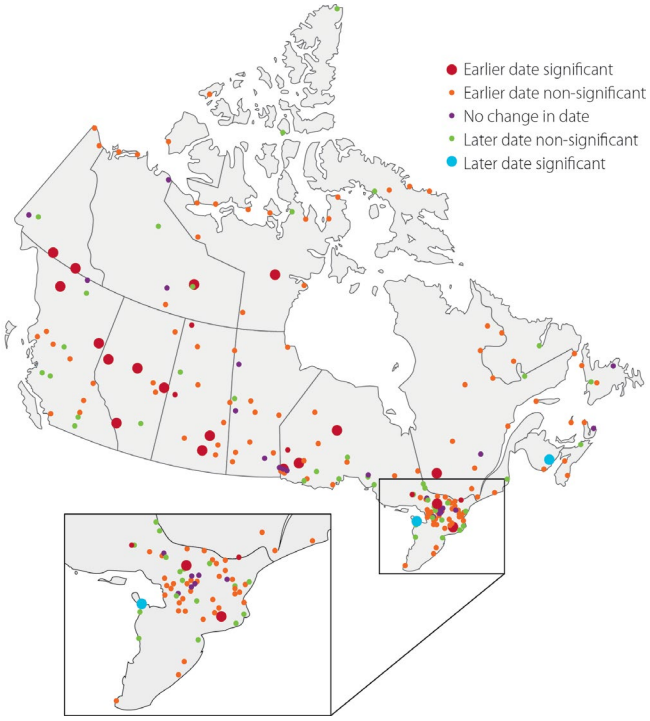

FIGURE 18: Changes in lake ice thaw dates across Canada, 1950-2005. Lakes with significant trends at the 90% confidence level and measurements covering at least 60% of the 1950 to 2005 time span *(Source: Icewatch, 2008)*.

Predicted changes in ice cover will impact the role of lakes on energy, water and biogeochemical processes in cold regions, and result in ecosystem changes within the lakes (Prowse et al., 2011). A shorter ice season and thinner freshwater ice will have detrimental impacts on the duration and stability of winter ice roads in northern Canada (Furgal and Prowse, 2008; *see also* Chapter 8).

3.5 SEA ICE

Numerous studies have documented the decline in Arctic summer sea ice cover that is evident over the satellite era (1979 to present), as well as the shift toward a younger, thinner ice cover (Stroeve et al., 2012a; Derksen et al., 2012 and references therein). The results reported on below highlight trends from recent studies of particular relevance to sea ice in northern Canadian waters (including the Arctic, Hudson Bay regions and northern east coast). Key indicators for tracking changes are ice extent at the end of summer (September) when sea ice coverage reaches its annual minimum, and extent at the end of winter (March) when coverage is at a maximum.

The decreasing trend in September Arctic sea ice extent, reported in the 2008 assessment (Lemmen et al., 2008) using data up to 2005, has continued. The rate of decline in the most recent decade (1999-2010) has been more rapid than in the earlier part of the record (1979-1998) (Stroeve et al., 2012a). The lowest September sea ice extents on record were all observed in the five summers from 2007-2011 (Maslanik et al., 2011; Perovich et al., 2011; Comiso, 2012). Sea ice extent reached a new record minimum in September 2012, at 18% below the previous record minimum set in 2007 (Perovich et al., 2012). Over the satellite sea ice observation era, end-of-summer sea ice has been declining by 13% per decade (from 1979-2012). A decreasing trend in maximum winter (March) sea ice extent has also become evident over the last decade, with the trend over the satellite era reported as -2.6% per decade (from 1979-2012) (Perovich et al., 2012) (Figure 19). The observed changes in sea ice extent have been attributed to a combination of forcing from increasing atmospheric greenhouse gas levels and natural variability in air temperatures, atmosphere/ocean circulation, and ocean temperature (Min et al., 2008; Stroeve et al., 2012a).

A number of studies have documented the shift from an ice cover formerly dominated by thick multi-year ice (MYI) to one increasingly dominated by thin first-year ice (FYI) (Derksen et al., 2012; Stroeve et al., 2012a; Comiso, 2012, Maslanik et al., 2011). Arctic Ocean (excluding the Canadian Arctic Archipelago region) MYI extent declined by 33% in March and by 50% in September between 1980-2011 (Maslanik et al., 2011). An observed increase in the variability of September sea ice cover since the early 1990s has been linked to this increasingly young ice (Stroeve et al., 2012a) as FYI can respond quickly to prevailing climate, melting rapidly during warm summers and then reforming and persisting when temperatures are sufficiently cool. Declines in Arctic sea ice extent and volume associated with ongoing climate warming are expected to continue over the coming decades. A nearly ice-free summer is now considered a strong possibility for the Arctic Ocean by the middle of the century, meaning that almost all of the ice would become FYI (although any remnant MYI would likely be found in the Canadian Archipelago region) (Meier et al., 2011; Massonnet, et al., 2012; Stroeve et al., 2012b; Wang and Overland, 2012).

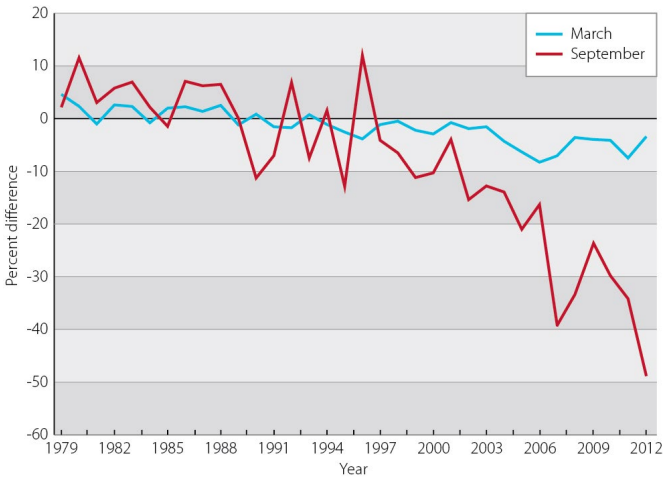

FIGURE 19: Trends in Arctic sea ice extent over the period 1979-2012 shown as time series of the percentage difference in ice extent in March and September relative to the mean values for the period 1979-2000. The rate of decrease for the March and September ice extents is -2.6% and -13% per decade, respectively (as determined by least squares linear regression). Both trends are statistically significant *(Source: Perovich et al., 2012)*.

A longer record of changing ice cover for Canadian sea ice regions (extending back to the 1960s) was developed using data from the historical Canadian Ice Service Digital Archive (CISDA) (Tivy et al., 2011; Howell et al., 2008a). These longer-term data sets confirm the decline of summer ice cover in northern Canadian waters, with total summer sea ice cover decreasing by 3 to 17% per decade (1968-2010), with the largest declines evident in the Hudson Strait and northern Labrador Sea (Figure 20). In almost all Canadian sea ice regions, the magnitude of the trend since 1968 is smaller than that since 1979 (Tivy et al., 2011), indicating a faster rate of decline in recent decades. The data also confirmed that substantial declines in summer MYI are evident in the southern Beaufort Sea and Foxe Basin (16% and 20% decreases per decade, respectively) (Figure 20). However, no statistically significant decline in MYI within the Canadian Arctic Archipelago (CAA) region is evident. The lack of significant declines in MYI in this area has been attributed to the inflow of MYI from the Arctic Ocean (Howell et al., 2009).

Ice conditions in the CAA are largely dependent on prevailing winds. Therefore, despite recent, consecutive, late summer 'open water' seasons in the Northwest Passage (NWP) (Perovich et al., 2011), ice conditions in the NWP, including the presence of MYI, are anticipated to remain variable and potentially hazardous for shipping for some time (Derksen et al., 2012; Howell et al., 2008a,b; Melling, 2002). Model projections of future ice conditions in the CAA support this conclusion (Sou and Flato, 2009; Stephenson et al., 2011). Projected declines in ice concentration and ice thickness within the CAA (A2 scenario) indicate that the southern route through the NWP could become consistently accessible to shipping by mid-century (using criteria for accessibility of 60% ice concentration and ice thickness less than 1.0 m). On the other hand, passage through the northern deep water route will remain limited by ice much of the time (~40%) (Sou and Flato, 2009). A similar conclusion of increased, but not consistently open accessibility of the NWP was reached by Stephenson et al. (2011).

Observations also indicate statistically significant declines in sea ice extent in the Labrador-Newfoundland region and in the Gulf of St. Lawrence (winter only; the Gulf is ice-free in summer) (Cavalieri and Parkinson, 2012; Hutchings et al., 2012), although there has been strong decadal-scale and interannual variability in southern areas (Colbourne et al., 2012; Galbraith et al., 2012b). Both CMIP3 and CMIP5 models indicate reductions in ice extent in the Northwest Atlantic over the coming decades, but the models generally do not resolve the details of observed ice patterns in the region. Considering the recent large declines in sea ice in areas such as the Gulf of St. Lawrence and Grand Bank, together with expected continued atmospheric warming, ice-free winters may occur within a couple of decades in warm years in such areas (Hammill and Galbraith, 2012), although partial ice cover will probably continue in colder years.

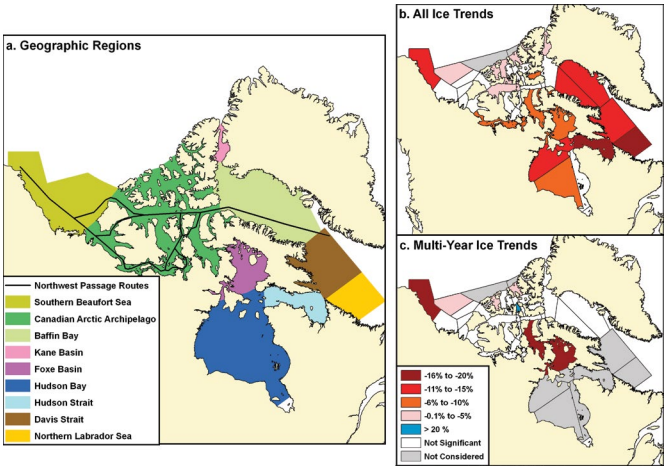

FIGURE 20: Trends in summer season total sea ice area (b) and multi-year sea ice area (c) for sea ice regions of Canada (a), from 1968-2010; units are % per decade. Only trends significant to the 95% confidence level are shown *(Source: Derksen et al., 2012, updated from Tivy et al., 2011).*

4. CHANGES TO FRESHWATER RESOURCES

Canada is considered to have an abundance of water resources with over 8500 rivers and 2 million lakes covering almost 9% of the total land area (Monk and Baird, 2011). Yet over three quarters of the volume of river flow is to the north, where the density of people and development is sparse.

A national perspective on freshwater availability is challenging, in large part due to great regional variations in climate and watershed characteristics. Long-term measurements at gauging stations with relatively undisturbed conditions are not evenly distributed across Canada. There are gaps in the Arctic Archipelago, southern Prairies, and highly developed portions of the country. Recent research on trends and climate change impacts has tended to be regionally focused. This review focuses on streamflow/runoff and lake levels as indicators of freshwater resources, drawing on long-term observational data and published research for these indicators.

4.1 OBSERVED CHANGES IN FRESHWATER AVAILABILITY

As described in the preceding sections, many key climate drivers of the hydrological regime are changing across Canada. For example, in many regions, the proportion of total precipitation falling as snow is declining; the spatial coverage and duration of snow cover are decreasing, affecting the timing and amount of spring runoff; and rising air temperatures are influencing evapotranspiration and water loss to the atmosphere. These trends have implications for the water balance and are manifested as changes in the amount and timing of water availability (Bates et al., 2008).

4.1.1 STREAMFLOW

Many gauging stations across Canada (Box 5) are exhibiting changes in streamflow/runoff with respect to median annual flow, median flow for selected months, annual maximum and minimum flow, timing and duration of events and variability of flows (Table 1). However, the patterns are not spatially consistent (Federal, Provincial and Territorial Governments of Canada, 2010; Monk et al., 2011). The winter months – December to February – had the highest proportion of gauging sites with statistically significant increases in runoff; April median runoff also increased. These trends may be related to warming in winter and spring. Runoff for May to September decreased with the greatest reduction in flow occurring in August (28% of stations showed a significant trend; Monk et al., 2011).

Trends in annual maximum flow, usually representative of the freshet associated with spring snowmelt and/or rain events, showed a widespread decrease (Figure 21A), although only 17% of stations had a statistically significant decreasing trend (Federal, Provincial and Territorial Governments of Canada, 2010). Annual minimum flow increased in the northwest and west, while in southern Canada, low flow decreased (Figure 21B). Similar annual and monthly trends have been reported in other studies (Khaliq et al., 2008; Abdul-Aziz and Burn, 2006; Burn et al., 2008; 2010; Cunderlik and Ouarda, 2009; Monk et al., 2011; 2012). Trend analysis results can be influenced by the length of the period analysed and natural variations of a decade or longer (Khaliq et al., 2008; Chen and Grasby, 2009).

BOX 5

STREAMFLOW ANALYSIS

Many results presented here are drawn from the Ecosystem Status and Trends (ESTR) National Assessment (Federal, Provincial and Territorial Governments of Canada, 2010; Monk and Baird, 2011) which developed a national picture of hydrologic trends for the period 1970-2005. Data from 172 streamflow gauging stations were abstracted from the Reference Hydrometric Basin Network (RHBN). These stations represent fairly "pristine" conditions where land use and land cover conditions are relatively stable and flow alterations from regulation, withdrawals or diversions are minimal (Brimley et al., 1999; Harvey et al., 1999).

4.1.2 LAKE LEVELS

Lake levels are highly visible indicators of changes in the water balance, human influences and availability of water in storage (Williamson et al., 2009). While there are many lakes throughout Canada, the limited number of lake observing sites precluded a national assessment of lake level trends. As such, this section discusses recent regional analyses of the Prairies and the Laurentian Great Lakes using water levels and/or net basin supply and its components as indicators.

Variable	Trend description
Magnitude of monthly median runoff	Few trends apparent. Strong SIT in April runoff, SIT in December, January, February and March runoff, and SDT for May to August runoff
Magnitude of minimum runoff (1-, 3-, 7-, 30-, and 90-day)	Majority NCT especially indicators with longer duration averaging period . Approximately one quarter of sites SDT
Magnitude of maximum runoff (1-, 3-, 7-, 30-, and 90-day)	Majority NCT but large number of sites (especially indicators with a long duration averaging period) showing TFDT
Timing of annual 1-day minimum	Few sites with significant trends Nearly half of sites showing TFIT towards (later) annual minimum
Timing of annual 1-day maximum	Few sites with significant trends. Majority of sites showing tendency towards (earlier) annual maximum
Frequency of extreme low flow events	Majority of sites NCT
Frequency of extreme high flow events	Majority of sites NCT
Duration of extreme events	Majority of sites NCT Slight trend towards SDT in duration of low pulse events
Flashiness of events	A few sites with significant trends for rise rate and fall rate. Tendency towards TFIT in fall rate and TFDT in rise rate for nearly half of sites. SIT in number of reversals for one third of sites

TABLE 1: Summary of national trends in streamflows based on the ESTR analysis for 1970 to 2005 and review of Canadian published literature on hydrologic trends *(Source: Monk et al. 2011; Monk and Baird, 2011)*. Note: SDT - significant declining trend ($p<0.1$); SIT - significant increasing trend ($p<0.1$); NCT - no clear trend or both increasing and decreasing trend; TFDT tendency for declining trend ($p>0.1$ i.e. not statistically significant) and TFIT tendency for increasing trend ($p>0.1$ i.e. not statistically significant).

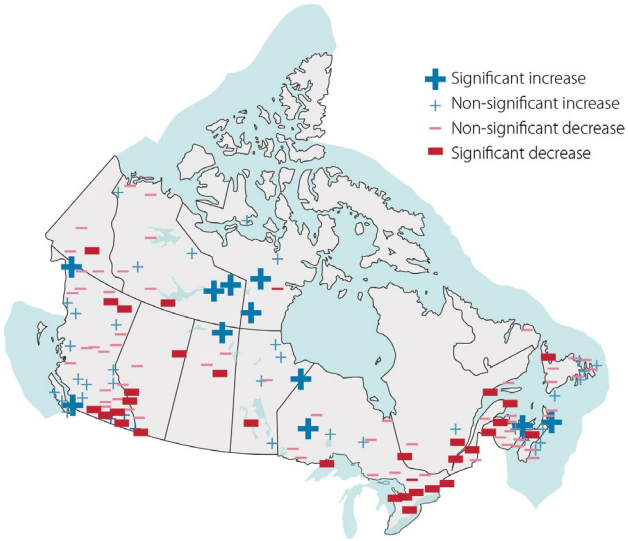

FIGURE 21 A: Trends in 1-day Maximum River Flow in 172 RHBN rivers (1970 to 2005) *(Source: Federal, Provincial and Territorial Governments of Canada, 2010).*

FIGURE 21 B: Trends in 1-day Minimum River Flow in 172 RHBN Rivers (1970 to 2005) *(Source: Federal, Provincial and Territorial Governments of Canada, 2010).*

THE PRAIRIES

Van der Kamp et al. (2008) developed a database of measured and reconstructed water levels for sixteen closed-basin lakes in the semi-arid Canadian Prairies to assess long-term water level patterns. Observations are not continuous over the period 1910-2006, but most are concentrated in the 1960s to present. Many lakes show a tendency of water level decline but this could reflect decadal variability. The decline is likely a result of the dynamic interplay between reduced runoff and precipitation inputs and an increase in losses due to evaporation. However, from the 1960s onward in east-central Saskatchewan, some lakes had rising water levels, likely reflecting climatological changes as well as other factors such as agricultural practices and land cover/land use changes.

LAURENTIAN GREAT LAKES

Water levels of the Laurentian Great Lakes fluctuate. Seasonally, water levels typically progress from a summer maximum to a minimum in winter/spring, with a documented earlier onset of the seasonal cycle and changes in the amplitude of water levels (Lenters, 2004; Argyilan and Forman, 2003). The lakes also exhibit interannual and inter-decadal fluctuations in water levels of less than 2.0 m (observed for the period 1918-2012), varying by lake (Wilcox et al., 2007; DFO 2013a). The most recent period of high water levels for all of the Great Lakes occurred in the 1980s. Levels subsequently declined rapidly, particularly from 1997-2000 (Assel et al., 2004). For the period 1998-2013, the upper Great Lakes – Superior, Michigan and Huron – have had a period of low water levels. Lake Michigan-Huron attained record low monthly mean levels in December 2012 and January 2013 (DFO, 2013b; Environment Canada, 2013) whereas Lake Superior set record low levels in August and September 2007 (DFO, 2007). A number of potential contributing factors have been identified, including climate change, vertical land movement and human modification of the system (e.g. dredging in the connecting channels) (International Upper Great Lakes Study – IUGLS, 2012). A similar water-level decline over this period in nearby Wisconsin seepage lakes suggests that regional climate variations may be a common driver (Stow et al., 2008). Water levels in the Great Lakes have also been correlated with large-scale atmosphere-ocean circulation patterns (e.g. Atlantic Multidecadal Oscillation (AMO), Pacific Decadal Oscillation (PDO) and El Niño-Southern Oscillation (ENSO) (Ghanbari and Bravo, 2008; Hanrahan et al., 2009; Wiles et al., 2009).

Water supply to the Great Lakes can also be represented by net basin supply (NBS) – a quantification of the factors, such as over-lake precipitation, runoff and evaporation that influence water levels. Analysis of these underlying factors and water-level trends in Lake Superior (1860-2007) and Lake Michigan-Huron (1860-2006) determined that there has been a negative linear trend in water levels – particularly since the end of the 20th century – that may have links to changes in evaporation and net precipitation (precipitation minus evaporation) (Sellinger et al., 2008; Lamon and Stow, 2010). The IUGLS (2012) undertook extensive trend and change-point analyses of water balance components for the period 1948 -2008 (Figure 22; *see also* Case Study 4 in Chapter 8). Evaporation has increased in all of the Great Lakes since 1948, but for most of the Lakes this has been offset by an increase in precipitation. This is not the case for the Lake Superior basin where evaporation has increased but precipitation has remained relatively constant, resulting in declining water supplies. Since most of the observed trends are within the range of natural variability, it is not possible to attribute these changes to climate change (Hayhoe et al., 2010).

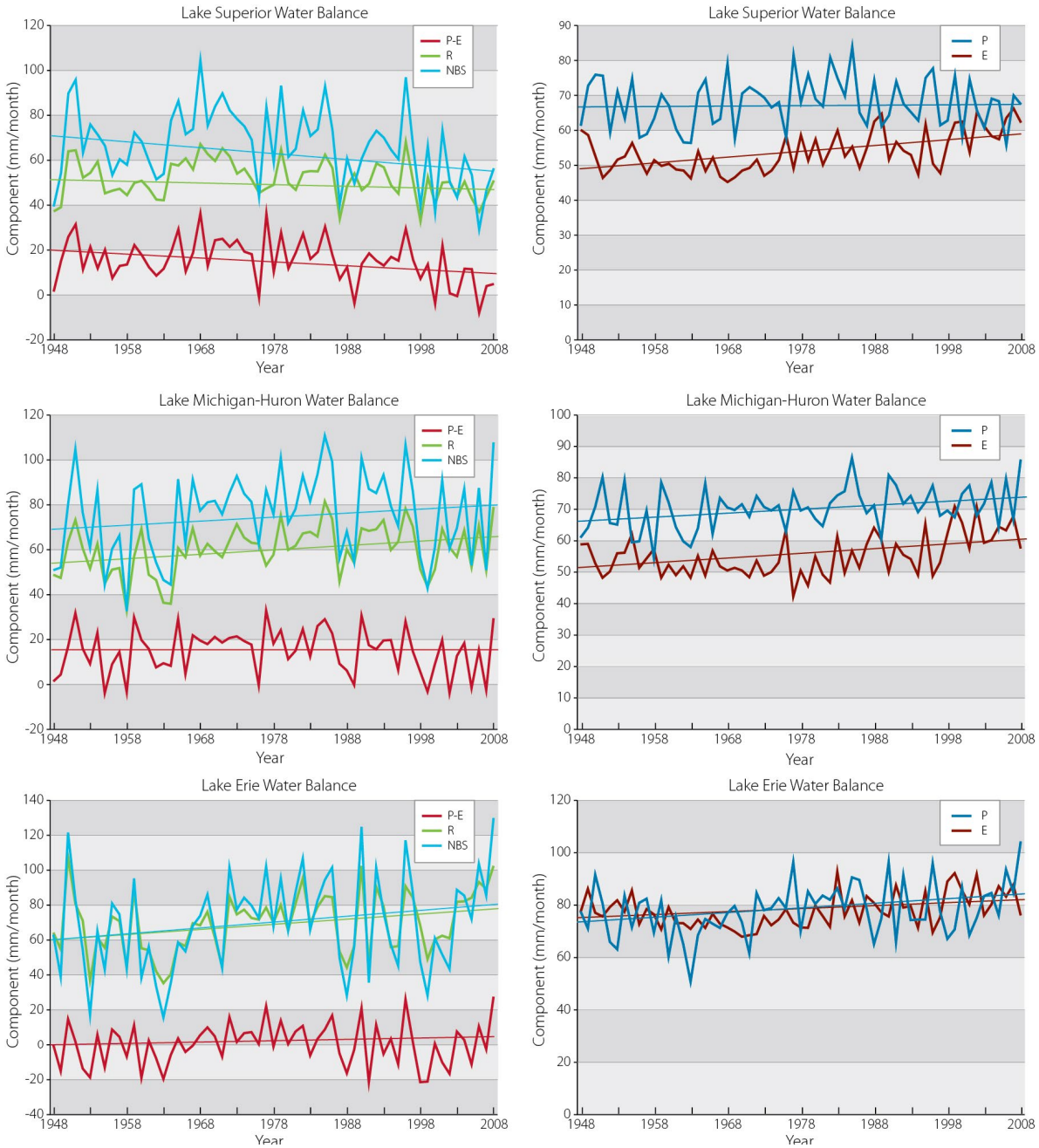

FIGURE 22: Trends in mean annual overlake precipitation (P), precipitation minus evaporation (P-E), runoff (R), and component net basin supply (NBS) for the upper Laurentian Great Lakes for the period 1946 to 2008. *(Source: IUGLS, 2012 derived from Fortin and Gronewold, 2012).*

4.2 PROJECTED CHANGES IN FRESHWATER AVAILABILITY

No national synthesis of studies investigating projected changes in surface water resources has been published since the review of studies within each regional chapter of Lemmen et al. (2008). However, several regional watershed-based studies have been published since 2008 that provide runoff projections for various climate change scenarios (Figure 23 and Table 2).

Region	Projections	Key References
Baker River, BC	Most scenarios for the 2050s project increased winter runoff and decreased summer runoff, including decreased snow water equivalent	Bennett et al., 2012
Campbell River, BC	Increase in 2050s winter runoff, and decrease in summer runoff; no consensus on changes in mean annual runoff	Schnorbus et al., 2011, 2012; Bennett et al., 2012
Trepanier Creek, Okanagan Basin, BC	Decrease in 2050s mean annual and summer streamflow, with spring freshet occurring 2 weeks earlier, compared with 1983-1993 period	Harma et al., 2012
Ingenika River, BC	Increase in 2050s winter runoff; no consensus on changes in summer runoff	Bennett et al., 2012
Fraser River, BC	No consensus for 2050s mean annual flow projection for the Fraser River; flow during summer would decline in all scenarios	Shrestha et al., 2012b
Athabasca River, AB	Decrease in 2080s mean annual flow, and annual minimum flow	Kerkhoven and Gan, 2011; Shrestha et al., 2012b
Southern Prairies (tributaries of Saskatchewan River, AB, SK)	Decreases in 2050s annual runoff, except for increase in Cline River, AB due to large increase in winter runoff	Lapp et al., 2009; Shepherd et al., 2010; Forbes et al., 2011; Kienzle et al., 2012; St. Jacques et al., 2012
Churchill, MB	In a hydrologic model intercomparison, 2 of 3 hydrologic model projections for a range of climate scenarios projected increases in annual runoff, while a 3rd model projected decreases	Bohrn, 2012
Lake Winnipeg – Upper Assiniboine and Morris Basins, MB	Increased annual runoff projected for the Upper Assiniboine, and for most scenarios in the Morris Basin	Shrestha et al., 2012a; Stantec, 2012
Spencer Creek, ON	Increase in mean annual and fall-winter streamflow, and decrease in March-April spring peak flow	Grillakis et al., 2011
Credit River, ON	Mixed projections of annual streamflow	EBNFLO Environmental and AquaResource Inc., 2010
Great Lakes, ON	Decrease in 2050s Net Basin Supply in Lakes Michigan-Huron and Erie; little change in Lake Superior	MacKay and Seglenieks, 2013; Chen et al., 2011
Tributaries of the St. Lawrence, QC (including Richelieu, St. François, Yamachiche, St. Maurice and Batiscan Rivers)	Increases in 2050s mean winter runoff, with most scenarios projecting decreased summer runoff and increased annual runoff	Boyer et al., 2010
Chaudière, QC	No consensus on projected changes in 2020s mean annual runoff	Quilbé et al., 2008
Pinus River Basin (Labrador, NL)	Increase in 2050s mean annual streamflow, with spring peak occurring 2 weeks earlier compared with the 1971-2000 period	Roberts et al., 2012

TABLE 2: Summary of projected changes in freshwater. *See also* Figure 23.

Most watersheds in Canada are influenced by snow accumulation and melt patterns. Maximum snow water equivalent (SWE) is projected to decline in coastal British Columbia, the Atlantic Provinces and the Great Lakes-St. Lawrence region, while increases are projected for the Arctic coast of Nunavut (Brown and Mote, 2009).

For watersheds that contain glaciers, glacier retreat has already been observed in British Columbia and Alberta (Stahl et al., 2008; Marshall et al., 2011; Jost et al., 2012), and this is projected to continue as the climate warms. As the ice melts, this is expected to influence runoff, particularly during summer. Marshall et al. (2011) assessed glacier runoff for

2000-2007 and future scenarios to 2100 (using SRES scenarios B1 and A1B) for Rocky Mountain glaciers contributing to the Bow, Red Deer, North Saskatchewan, Athabasca and Peace Rivers. Projected changes in glacier volume range from -80% (Athabasca) to -100% (Red Deer). Projected glacier runoff changes between 2000 and 2050 for the A1B scenario are -80% for North Saskatchewan River, -100% for Bow and Red Deer Rivers, -75% for Peace River, and -60% for Athabasca River. As glacier runoff contributed around 7% to summer runoff in the Bow and North Saskatchewan Rivers in 2000-2007, projected reductions need to be accounted for in projections of streamflow during low flow periods in summer and fall. Studies are also available for drainage basins with glaciers in British Columbia (Bürger et al., 2011; Stahl et al., 2008; Jost et al., 2012).

Future lake levels in the Great Lakes basin, for three time periods in the 21st century (relative to 1970 to 1999), were evaluated recently by Angel and Kunkel (2010) using more than 500 scenarios based on Global Climate Model (GCM) runs forced by the B1, A1B, and A2 emission scenarios. While the majority of simulations project water level declines, higher water levels are also a possibility. For the 2050-2064 period, a wide range in lake levels for Lake Michigan-Huron were projected, ranging from a decline of around 1.5 m to an increase of more than 1 m. Based on the research available, including recent RCM results (e.g. MacKay and Seglenieks, 2013) the IUGLS concluded that in the short term, water level reductions may not be as extreme as projected in earlier climate change assessments, and while future lower water levels are likely, the possibility of higher levels must also be incorporated in water management and planning (IUGLS, 2012).

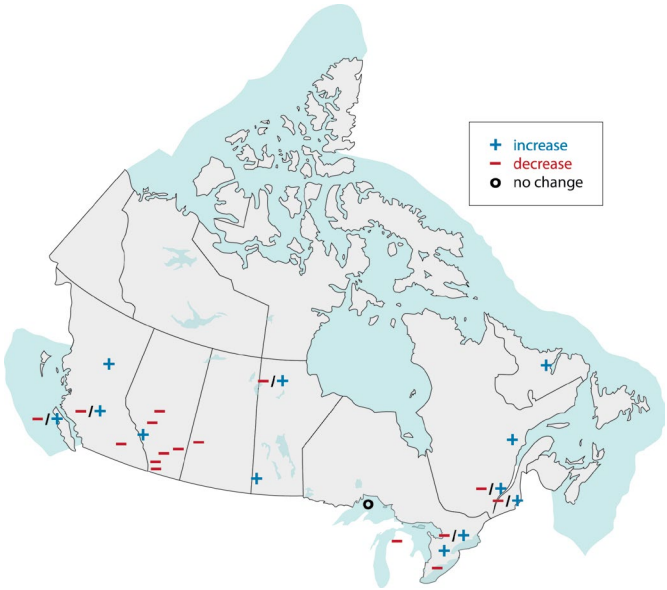

FIGURE 23: Projected changes in annual runoff for the 2050s period, from research published during 2008-2013. A '-/+' means that for a particular study location, scenario projections include both decreases and increases. *See* Table 2 for study references.

5. CHANGES IN OCEAN CLIMATE

Canada is bounded by three oceans with continental shelves, and has a long coastline with many embayments, straits, estuaries and coastal seas. Relevant ocean climate changes were discussed in the regional chapters of Lemmen et al. (2008) and are briefly discussed here as background to the thematic chapters in this updated report.

The primary observed large-scale ocean climate changes are summarized schematically in Figure 24. They include widespread warming and increasing carbon dioxide (CO_2) in the upper ocean (contributing to sea level rise and decreasing pH), reduced sea ice extent and freshening at high latitudes, and increasing surface salinity at low latitudes.

5.1 OCEAN TEMPERATURE

Influences of climate change on ocean temperature, salinity, acidity, sea level and other variables are evident from observational datasets around Canada (Hutchings et al., 2012; Christian and Foreman, 2013; Loder et al., 2013a; Steiner et al., 2013). However, it remains challenging to distinguish between anthropogenic and natural variability in the relatively short available time series for most variables. There is strong natural decadal-scale atmospheric and oceanographic variability (such as the Pacific Decadal, Arctic, North Atlantic and Atlantic Multi-decadal Oscillations) in various regions (*see* Box 3). In turn, this influences important ocean features such as El Niño in the Pacific, and the Labrador Current in the Atlantic, on regional scales.

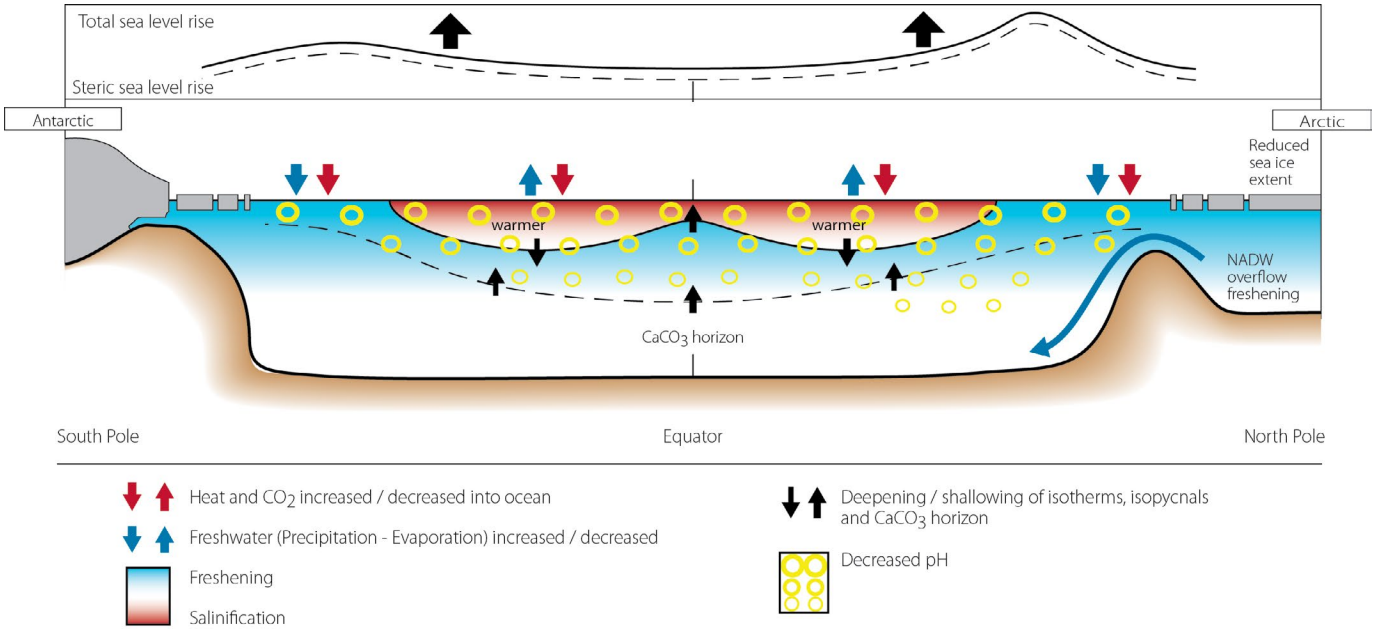

FIGURE 24: Schematic side view of prominent observed climate changes in the ocean. The legend identifies the direction of the changes. The "total" sea level rise is referred to as the "absolute" sea level rise, and the $CaCO_3$ (calcium carbonate) "horizon" as the "saturation depth" in the text *(Source: Bindoff et al., 2007).*

5.1.1 PACIFIC COAST

Long-term warming trends of about 0.1°C per decade are present in coastal ocean temperature observations taken at British Columbia lighthouses over the past 75 years, and in offshore upper-ocean (10 to 50 m below surface) observations on Line P[3] in the Northeast Pacific over the past 55 years (Figure 25; Irvine and Crawford, 2012). Upper-ocean temperatures off British Columbia show strong natural variability associated with El Niño, La Niña and the PDO. In particular, temperatures in recent years have been cooler than in the previous two decades due to a Pacific-wide weather pattern associated with La Niña conditions in these years. Subsurface (100 to 150 m) waters on Line P show a weaker warming trend (~0.05°C per decade) and a decadal-scale variation resembling that in the upper-ocean waters. These warming trends are qualitatively consistent with global analyses of sea surface (Yasunaka and Hanawa, 2011) and subsurface (Bindoff et al., 2007) temperature datasets that indicate an overall ocean warming trend in the Northeast Pacific over the past century.

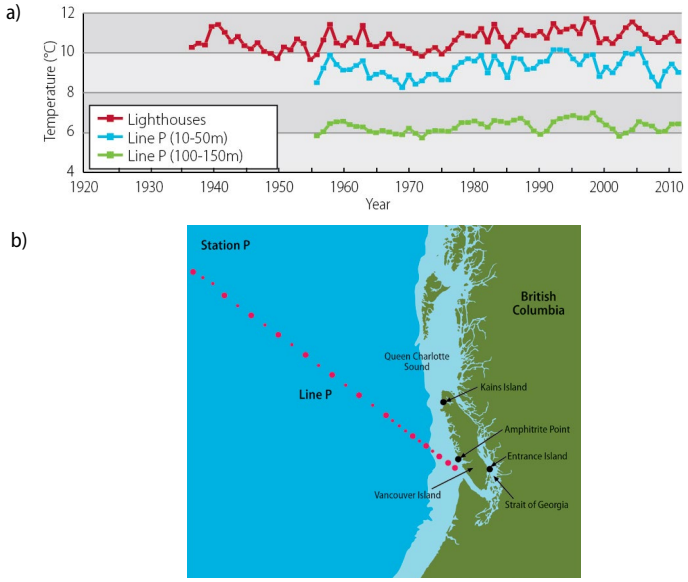

FIGURE 25: **a)** Annual-mean indices of ocean temperature in the Pacific Ocean off British Columbia, from DFO monitoring programs reported in Irvine and Crawford (2012). The Lighthouses index is the average of surface temperature measured daily at Amphitrite Point, Entrance Island and Kains Island (Chandler, 2012), while the Line P indices are based on Figure 12 of Robert et al. (2012). The trends (0.1, 0.09 and 0.05°C per decade, respectively, for the three time series), are all significantly different from zero at the 95% confidence level. **b)** Location map for Pacific Ocean observation sites.

[3] Line P is a Fisheries and Oceans Canada (DFO) monitoring line of 26 oceanographic stations extending about 1400 km offshore from the southwest coast of Vancouver Island to the former position of Ocean Weather Station (OWS) Papa in the Gulf of Alaska (Crawford et al. 2007; Figure 25b).

5.1.2 ATLANTIC COAST

Off Atlantic Canada, upper-ocean temperature observations over the past 60 to 80 years from mid-latitude regions to the west of the Grand Bank (such as the Bay of Fundy; Figure 26) generally show warming trends similar in magnitude to those in the Pacific. Warming trends are apparent in both the surface and near-bottom waters in the Gulf of St. Lawrence (Galbraith et al., 2012a) and Scotian Shelf (Hebert et al., 2012). The surface warming in the Gulf is consistent with increasing air temperatures (Galbraith et al., 2012b) and the higher near-bottom warming rate there is related to an increasing influence of subtropical waters from the Gulf Stream (Gilbert et al., 2005).

In contrast, there has been no significant warming trend over the past 60 to 80 years in the upper 150 m of the Labrador Sea and Newfoundland Shelf (Figure 26), where decadal-scale natural variability associated with the NAO and AMO has been dominant (Yashayaev, 2007; ICES, 2011; Loder et al., 2013b). The absence of a long-term trend in this region is consistent with the large area south of Greenland where there has been no net warming observed in surface air and water temperatures over the past century (Trenberth et al., 2007; Yasunaka and Hanawa, 2011). Over the past 2 to 3 decades there has been ocean warming in the Labrador-Newfoundland region.

5.1.3 ARCTIC COAST

There are no long time-series of ocean temperature in the Canadian Arctic, but the clear signals of warming in air temperature and sea ice observations, together with available ocean observations and model simulations (Galbraith and Larouche, 2011; Timmermans, 2012), point to the occurrence of upper-ocean warming in most areas.

5.1.4 PROJECTIONS

Projections from the CMIP3 and CMIP5 (*see* Box 4) models generally indicate widespread warming of the upper ocean around Canada during the 21st century, with substantial seasonal and spatial variability (Meehl et al., 2007b; Capotondi et al., 2012). Projected surface temperature increases from 1951-2000 to 2051-2100 from CMIP3 models for the A1B scenario are generally in the 1 to 3°C range (Capotondi et al., 2012). One exception is the northern North Atlantic south of Greenland, where most models indicate more limited warming, apparently associated with a reduction in the northward ocean transport of heat by the Atlantic Meridional Overturning Circulation (AMOC) (Drijfhout et al., 2012; Hutchings et al., 2012). However, the extent to which this projected ocean temperature anomaly will extend into the Labrador and Newfoundland coastal waters is uncertain, in view of the difficulties that global models have in resolving ice-ocean variability in the Labrador Sea (de Jong et al., 2009).

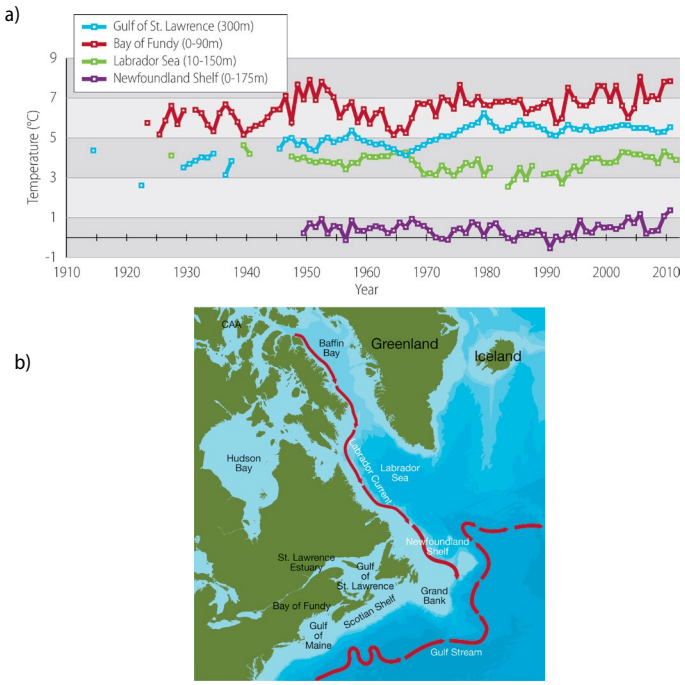

FIGURE 26: a) Annual-mean indices of ocean temperature off Atlantic Canada from DFO monitoring programs. The Bay of Fundy (BF) time series (Hebert et al. 2012) are from the Prince 5 station in Passamaquoddy Bay which has been sampled regularly since 1926, while the Newfoundland Shelf (NS) time series (Colbourne et al., 2012) are from Station 27 off St. John's which has been sampled in most months since 1950. The Gulf of St. Lawrence (GSL) time series is for near-bottom waters (Galbraith et al., 2012a) and the central Labrador Sea (LS) time series is from the vicinity of former OWS Bravo (Yashayaev and Greenan 2012), with both based on all available data. The trends for the BF and GSL time series (0.14 and 0.22°C per decade, respectively) are significant at the 95% confidence level, but those from the NS and LS series are not significant. **b)** Location map for Atlantic Ocean observation sites.

5.2 OCEAN SALINITY AND DENSITY STRATIFICATION

5.2.1 SALINITY

Ocean salinity is an important contributor to ocean climate and marine ecosystems since together with temperature and pressure (depth), it determines the density of seawater, which in turn affects ocean circulation, vertical density stratification, and vertical mixing. Changes in salinity occur in response to precipitation, evaporation, freshwater run-off from the continent, melting and freezing of sea ice, and ocean circulation and mixing, such that there is more spatial variability in salinity than temperature. Away from areas with sea ice and continental run-off, changes in ocean salinity can also be used to infer changes in precipitation-minus-

evaporation (and hence the hydrological cycle) over the ocean (e.g. Helm et al., 2010).

Salinity in most locations off Pacific and Atlantic Canada over the past 60 to 80 years has been strongly influenced by decadal-scale variability similar to that observed for ocean temperature (Petrie, 2007; Yashayaev, 2007; Irvine and Crawford, 2012). Long-term trends are weak (<0.1 psu[4] per decade; Figures 27 and 28). There are also distinct local variations in some areas, apparently related to river run-off. There are indications of a long-term decrease in salinity in the near-surface waters over the Pacific shelf and offshore, but there are indications of increases in salinity in the Strait of Georgia and at the 150 m depth at Station P (Figure 27; Chandler, 2012; Freeland, 2013). The decreases in offshore surface values are consistent with the large-scale decrease observed in the Northeast Pacific (Durack and Wijffels, 2010).

On the Atlantic coast, long-term salinity changes have also been weak compared to decadal-scale variability, and have varied with location and with depth in some areas (Figure 28; ICES 2011). On the Scotian Shelf and in the Gulf of St. Lawrence and Bay of Fundy, there has been an overall tendency towards decreasing salinity in the near-surface waters but there has been increasing salinity in the deep near-bottom waters consistent with the poleward creep of subtropical waters (Gilbert et al., 2005; Hebert et al., 2012; Wu et al., 2012). Salinity variability off Labrador and Newfoundland has also been dominated by decadal-scale variability (Yashayaev, 2007; Colbourne et al., 2012), with no net change over 80 years in the Labrador Sea and a weak negative trend on the Newfoundland Shelf. More limited observations from the Arctic indicate freshening in most areas, but increased salinity in some others (e.g. Timmermans, 2012).

With the intensified hydrological cycle and land and sea ice melting expected in northern latitudes, long-term salinity decreases are generally expected in mid-to-high latitude waters such as those off Canada (Meehl et al., 2007b; Capotondi et al., 2012). Projected surface salinity decreases from 1951-2000 to 2051-2100 from CMIP3 models for the A1B scenario are generally in the range of -1 to -0.4 psu (Capotondi et al., 2012). An exception is the offshore slope and deep shelf waters between the Gulfs of St. Lawrence and Maine which are projected to have increased salinity by up to +0.4 psu because of the poleward creep of the North Atlantic's subtropical gyre and increased subtropical evaporation.

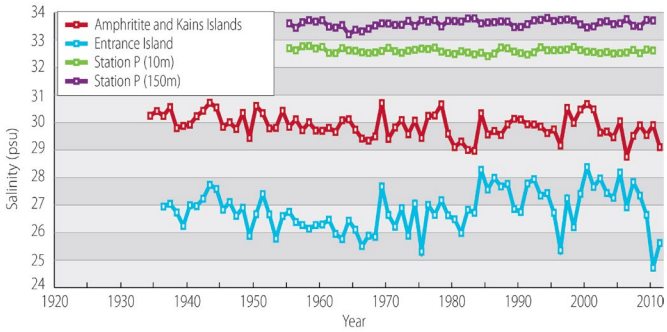

FIGURE 27: Annual-mean indices of ocean salinity in the Pacific Ocean off British Columbia, from DFO monitoring programs reported in Irvine and Crawford (2012). Values are in practical salinity units (psu), or parts per thousand. The Lighthouse time series (Amphitrite Point, Kains and Entrance Islands) are from Chandler (2012), while the Station P time series from the site of former OWS Papa are updates of those in Whitney et al. (2007). The trends for Station P (-0.01 and +0.02 psu per decade for 10 m and 150 m, respectively), and for Amphitrite Point and Kains Island (-0.06 psu per decade) are significant at the 95% confidence level, while the trend for Entrance Island (+0.07 per decade) is significant at the 93% confidence level. *See* Figure 25b for site locations.

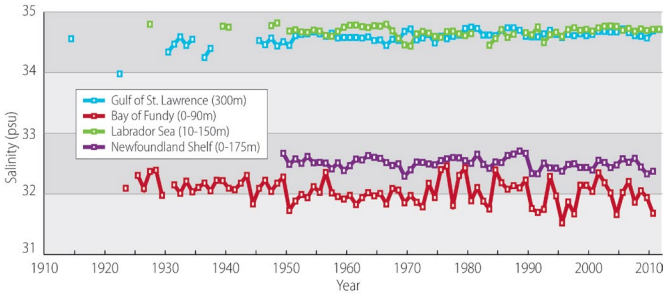

FIGURE 28: Annual-mean indices of ocean salinity off Atlantic Canada from DFO monitoring programs. The data sources and sites are the same as those for temperature in Figure 26. The decreasing trend for the BF (-0.02 psu per decade) and the increasing trend at depth in the GSL (+0.03 psu per decade) are significant at the 95% confidence level. The decreasing trend for the NS (-0.01 psu per decade) is significant at the 94% level, and the trend for the LS is not significant (*see* Figure 26b for site locations).

[4] Practical Salinity Unit, which is a unit based on the properties of sea water conductivity. It corresponds to parts per thousand (ppt) or to grams of salt per kilogram of water (g/kg).

5.2.2 VERTICAL DENSITY STRATIFICATION

The seasonally varying vertical stratification in upper-ocean water density is very important to ocean biogeochemistry and marine ecosystems (Box 6). Together with wind and tidal energy, it influences the vertical exchange of important dissolved and suspended materials, influencing atmospheric ventilation (e.g. CO_2 and oxygen) of subsurface waters, the upward supply of nutrients to the near-surface waters where phytoplankton grow, the suspension of phytoplankton, and the sinking rate of particulate material to greater depths.

BOX 6

OCEAN DENSITY STRATIFICATION

The density of seawater varies with its temperature, salinity, and depth. Ocean density stratification refers to the vertical gradient in water density whereby light, relatively warm and generally fresh near-surface water overlies cold, denser subsurface water. Important seasonal stratification develops in the upper ocean in spring and summer, as a result of the warming of near-surface water by solar radiation and atmospheric heating, and of near-surface freshening due to continental run-off. Strong stratification (large density gradients) tends to reduce vertical mixing in the ocean. Thus, the spatial and temporal variability of the stratification has important implications for mixing heat and CO_2 down into the ocean and for mixing nutrients (for plankton growth) up into the surface layers.

Global warming is resulting in increased upper-ocean vertical stratification in most ocean regions, due to the warming and hence lightening of surface waters. In most of the waters around Canada, this is expected to be reinforced by the freshening (and hence additional lightening) of near-surface waters (Capotondi et al., 2012). The observed long-term temperature and salinity trends off Pacific and Atlantic Canada are contributing to a long-term increase in stratification in many areas (Colbourne et al., 2012; Hebert et al., 2012; Freeland, 2013), although these changes are still dominated in some regions by decadal-scale variability (Figures 25-28). CMIP3 model projections (Capotondi et al., 2012) indicate increased upper-ocean stratification during this century around all of Canada, with both temperature and salinity contributing in most areas except in the subpolar North Atlantic where the freshening effect dominates and the subtropical North Atlantic where the warming effect dominates. This increased stratification will have important consequences for other ocean properties through reduced vertical mixing and ventilation of subsurface waters (e.g. Helm et al., 2011), and a general reduced supply of nutrients to the near-surface waters (e.g. Hutchings et al., 2012).

5.3 OCEAN HYPOXIA AND ACIDITY

5.3.1 OCEAN HYPOXIA

Observations off both Pacific and Atlantic Canada are indicating a general decline in the concentration of dissolved oxygen in subsurface (100 to 400 m) waters (Figure 29) below the more continually ventilated surface layer (Gilbert et al., 2005; Whitney et al., 2007; Hutchings et al., 2012; Crawford and Peña, 2013). This can be attributed to a combination of increasing temperatures (hence reduced solubility of oxygen) and increasing upper-ocean stratification (hence reduced ventilation), the poleward creep of subtropical waters in the Atlantic, and eutrophication from river run-off and biological productivity in some coastal areas (Gilbert et al., 2010). Concentrations in some areas are at or approaching "hypoxic" conditions (*see* Box 7), which are detrimental to marine organisms (e.g. Mucci et al., 2011; Bianucci and Denman, 2011), in part related to these climate change effects.

BOX 7

AQUATIC AND OCEAN HYPOXIA

The availability of dissolved oxygen is important to aquatic life. Ocean "hypoxia" is generally considered to occur when dissolved oxygen concentrations are lower than 60 to 80 µmole/kg. Aquatic hypoxia is best known in some lakes and coastal zones where oxidation of organic matter from runoff and plankton growth can result in local oxygen depletion and adverse conditions for aquatic life. The increasing temperature and increasing density stratification (and reduced ventilation) associated with climate change are acting to reduce dissolved oxygen levels in subsurface waters.

Low oxygen concentrations are already a serious concern for ecosystems and fisheries off Canada's Pacific coast where concentrations are naturally hypoxic at depths of 400 to 1000 m (Station P in Figure 29; Whitney et al., 2007; *see also* Chapter 4 – Food Production). This is, in part, due to these waters not having been in contact with the atmosphere for centuries during which time they have become oxygen-depleted by the oxidation of sinking organic matter. Intermittent upwelling of these waters onto the Pacific Shelf has resulted in the increasing occurrence of hypoxic bottom water conditions (Figure 29; Crawford and Peña 2013), particularly in summer, with natural decadal-scale variability clearly apparent in the observations.

Low oxygen concentrations have also become a serious issue at depth in the St. Lawrence Estuary (Figure 29), primarily due to increasing subtropical water at depth (Gilbert et al., 2005). More limited observations indicate that dissolved oxygen is also decreasing at depth on the Scotian Shelf (Petrie and Yeats, 2000) and in the Labrador Sea (Greenan et al., 2010). Past and future changes in dissolved oxygen in the Arctic are uncertain because of multiple (and in some cases offsetting) factors. For example, reduced ice cover can have different influences on stratification and ventilation, as well as on biological processes (Gilbert et al., 2010). The observed trends of reduced subsurface oxygen levels off Canada's other coasts are expected to continue with increasing CO_2 and heat in the atmosphere, and with increasing upper-ocean stratification in most areas (Meehl et al., 2007b; Hutchings et al., 2012). The effects of climate change are of particular concern in areas with or approaching hypoxic conditions.

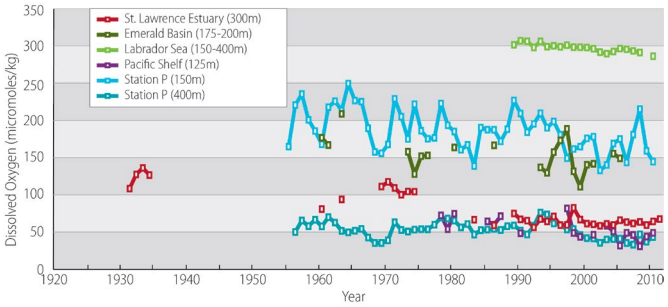

FIGURE 29: Annual indices of dissolved oxygen concentrations in subsurface waters off Pacific and Atlantic Canada, from DFO monitoring and other observation programs. The Pacific indices from Station P are updates of those in Whitney et al. (2007), while those for the Pacific Shelf are from Crawford and Peña (2013). The indices from the St. Lawrence Estuary, Emerald Basin (Scotian Shelf) and Labrador Sea are updates of those from Gilbert et al. (2005), Petrie and Yeats (2000) and Greenan et al. (2010). All trends are significant at the 95% confidence level: those for the first 5 sites listed are in the range of -7 to -9 µmole/kg per decade, and that at Station P (400 m) is -2 µmole/kg per decade. *See* Figures 25b and 26b for site locations.

5.3.2 OCEAN ACIDITY

Changes in ocean acidification (*see* Box 8) have a number of adverse implications for marine ecosystems, including a reduction in the stability of the carbonate ions used by marine organisms to build shells and skeletal structures (Doney et al. 2009; Hutchings et al., 2012).

BOX 8

OCEAN ACIDIFICATION

The increasing input of anthropogenic CO_2 from the atmosphere to the ocean is increasing ocean acidity. CO_2 reacts with seawater to generate hydrogen ions and form carbonic acid, which makes seawater more acidic and lowers its pH. Ocean acidification is therefore a direct consequence of CO_2 emissions. The IPCC Fourth Assessment (AR4) reported that ocean pH has decreased by 0.1 units since 1750 as a result of the uptake of CO_2 from the atmosphere, which represents a 30% increase in acidity.

Increased acidity reduces the concentration of carbonate ions in the ocean. The carbonate ion is used by many organisms to build shells or skeletons. The so-called "saturation depths" or "horizons", below which various carbonate minerals such as aragonite and calcite dissolve more readily than they can be formed, are becoming shallower. This limits the areas of the ocean that are suitable habitats for many marine organisms.

Observations in the Arctic and Atlantic waters off Canada (Yamamoto-Kawai et al., 2009; Greenan et al., 2010) indicate decreasing pH at rates similar to those observed globally; namely, by ~0.1 pH units since pre-industrial times (Bindoff et al., 2007). Acidification is most pronounced in cold fresh Arctic waters where carbonate saturation depths are already shallow (Yamamoto-Kawai et al., 2009; Azetsu-Scott et al. 2010), and at depth in the St. Lawrence Estuary where increasing subtropical influences and biological processes have resulted in the pH decrease being a factor of 4 to 6 times larger than that in the global surface waters (Mucci et al., 2011). Widespread ocean acidification has also been detected in the North Pacific Ocean (Feely et al., 2008). In some areas off all three coasts, waters are already considered to be "corrosive" to some calcareous organisms, i.e. capable of dissolving their shells and skeletons (e.g. Feely et al., 2008; Yamamoto-Kawai et al., 2009; Mucci et al. 2011).

Projections based on the IPCC SRES scenarios (*see* Box 4) indicate further global reductions in pH of between 0.14 and 0.35 units in the 21st century (Feely et al. 2009; Hutchings et al., 2012), but larger reductions may occur in local areas. The effects will be particularly significant in the high-latitude waters around Canada where the aragonite saturation depth could shallow into the upper 50 m by 2100 (Denman et al. 2011).

5.4 SEA LEVEL CHANGE

5.4.1 PAST AND PRESENT MEAN SEA LEVEL CHANGE

Global mean sea level rose about 21 cm between 1880 and 2012 (Figure 30). The rate of rise (Box 9) increased between the 20th and early 21st centuries. In the 20th century, sea level rose 1.7 ± 0.5 mm/yr, corresponding to about 17 cm over the course of the century while from 1993 to 2003, it rose 3.1 ± 0.7 mm/yr (Bindoff et al., 2007). It has continued at a similar rate to 2009 (Nerem et al., 2010) and in recent years (*see* Figure 30). The sources of sea-level rise include thermal expansion of the upper ocean and melt-water from glaciers, ice caps, and the Greenland and Antarctic ice sheets. Sea level rise is not uniform across the various oceans. Observed global sea-level change has exhibited substantial spatial variability, even over several decades (Meyssignac et al., 2012), mainly due to long-term spatial variability in thermal expansion and changes to salinity. Other effects, such as uneven melt-water redistribution, also contribute to spatial variability.

Much of the Canadian land mass is experiencing uplift due to glacial isostatic adjustment, which is the delayed rebounding of the land surface in response to the removal of the weight of the continental ice sheets during their retreat at the end of the last ice age (Figure 31; Peltier, 2004). The coastlines of Hudson Bay and the central Arctic Archipelago are rising rapidly and have been for thousands of years due to this adjustment, causing sea level to fall. At Churchill, Manitoba, the tide gauge shows sea-level fall of nearly 10 mm/yr since 1940 (Wolf et al., 2006), consistent with a measurement of crustal uplift slightly in excess of 10 mm/yr (Mazzotti et al., 2011).

BOX 9

ABSOLUTE AND RELATIVE SEA LEVEL CHANGES

Global sea-level change is commonly discussed in terms of "absolute" sea level, meaning that it is referenced to the centre of the Earth. At coastal locations, the sea-level change that is observed or experienced relative to a fixed location on land is known as relative sea-level change. Relative sea-level change is the result of absolute sea-level change and vertical land motion, both of which can vary from one location to another. Land uplift decreases relative sea-level rise and land subsidence increases it. In determining relative sea-level changes across Canada, vertical land motion (uplift and subsidence) plays a predominant role, although regional variations in absolute sea-level change are also important.

Outside the area of uplift, the land is sinking at lower rates. The sinking of land is due to the slow flow of rock deep in the Earth's mantle from subsiding regions towards uplifting regions. This reverses the process of flow away from regions depressed by ice sheets in the past. Most of the Maritimes, much of Newfoundland, the Yukon coast, the mainland coast of the Northwest Territories and some of its islands, and the

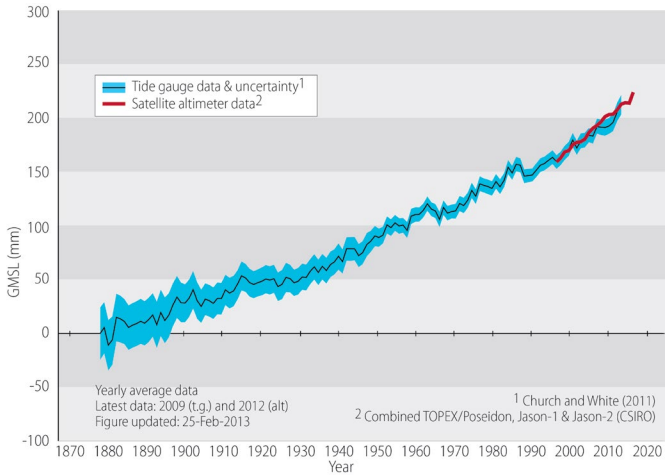

FIGURE 30: Observed global mean sea-level from 1880 to 2012 (*Source: Commonwealth Scientific and Industrial Research Organization [CSIRO], www.cmar.csiro.au/sealevel/ accessed June 17, 2013*). The observations are based on tide gauge data (1880 to 2009) and TOPEX/Poseidon, Jason-1, and Jason-2 satellite altimetry (1993-2012).

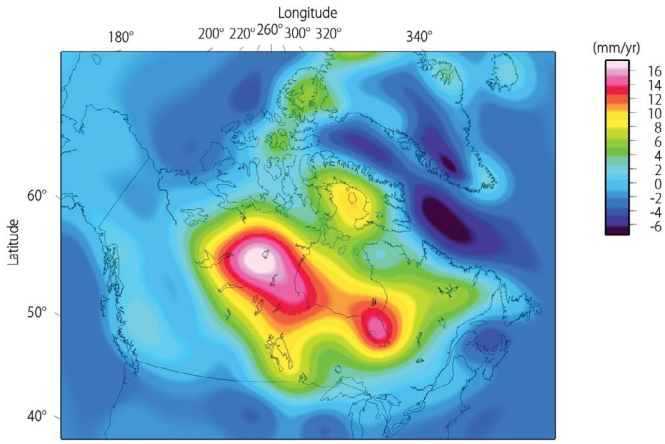

FIGURE 31: Present-day vertical crustal motion (in millimetres per year) predicted by the ICE-5G model of glacial isostatic adjustment (*Source: Peltier, 2004*). Relative sea level is presently falling in regions where the land is rising rapidly, such as Hudson Bay. Areas that are sinking, such as most of the Maritimes, experience relative sea-level rise that is larger than the global value. The model predictions do not include the significant vertical crustal motion in coastal British Columbia caused by active tectonics.

east coast of Baffin Island in Nunavut, are subsiding. These regions have experienced relative sea-level rise over the past few thousand years. At Halifax and Charlottetown, tide-gauge records show that relative sea level has risen at about 3.2 mm/yr throughout most of the 20th century (Forbes et al., 2004, 2009), nearly double the 20th century value of global sea-level rise. At Tuktoyaktuk, on the Beaufort Coast of the Northwest Territories, relative sea-level has risen at 3.5 mm/yr in the past half-century, consistent with a combination of global sea-level rise and local land subsidence (Forbes et al., 2010).

Relative sea-level rise in British Columbia has generally been smaller than in the Maritimes, with differences along the coastline largely arising from vertical land motion due to movement of tectonic plates offshore. The effects of past and present-day mass fluctuations of mountain glaciers and a residual glacial isostatic adjustment effect from the last continental glaciation are also present. In the 20th and early 21st century (1909 to 2006), sea level rose at an average rate of 0.6 mm/yr in Vancouver and Victoria, and 1.3 mm/yr in Prince Rupert, and fell by 0.9 mm/yr in Tofino (Mazzotti et al., 2008).

Another geological factor that contributes to relative sea-level change is sediment compaction. On the Fraser Delta, ongoing subsidence due to sediment compaction has been measured at 1-2 mm/yr (Mazzotti et al., 2008, 2009). Similarly, measurements on the Mackenzie Delta in the Northwest Territories show subsidence of up to several millimetres per year relative to a nearby stable reference point. The additional subsidence of the Delta further contributes to sea-level rise on this isostatically subsiding shoreline (Forbes et al., 2010).

5.4.2 FUTURE CHANGES TO MEAN SEA LEVEL

Global mean sea-level will continue to rise in the 21st century (Figure 32), but there is uncertainty regarding the rate. As projected by the IPCC AR4, the increase in global sea level over the 21st century, relative to the last two decades of the 20th century, would range from 18 to 59 cm depending on the emission scenario (Meehl et al., 2007b). For all scenarios, the thermal expansion component dominated, representing 70 to 75% of the central estimates of the sea level rise by the end of the century (Meehl et al., 2007b). The report also considered that an additional sea-level rise of 10 to 20 cm from accelerated glacier discharge to the oceans could be possible. These results were obtained from process-based models incorporating physical laws and known properties of the atmosphere, oceans, glaciers and ice sheets. Updated projections following the IPCC approach (e.g. Church et al., 2011) indicate a global sea-level rise of about 20 to 80 cm by 2100.

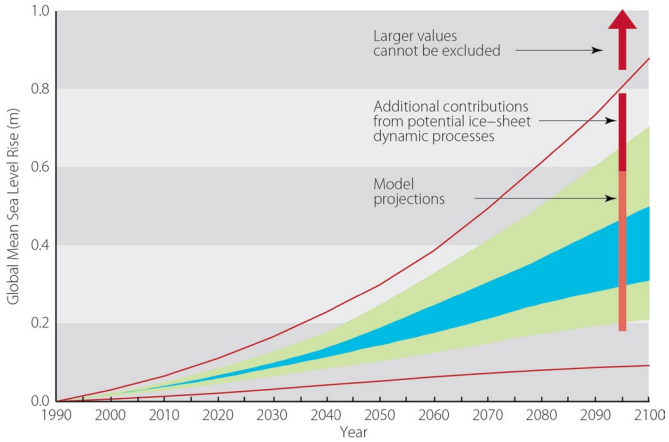

FIGURE 32: Projected global sea-level rise for the 21st century from the IPCC Third Assessment Report (TAR; blue and green shading and red lines; IPCC, 2001) and at the end of the century from the Fourth Assessment Report (AR4; coloured bars; IPCC, 2007). For the TAR, the blue shading shows the variation in the mean projections for a range of emissions scenarios, the green shading shows the range of all model projections, and the outer lines indicate an additional uncertainty from land ice. For AR4, the light red bar shows the range of model predictions, the dark red bar indicates a possible additional contribution from Greenland and Antarctic ice-sheet dynamics, and the dark red arrow shows that larger amounts of sea-level rise cannot be excluded[5] *(modified from Church et al., 2008)*.

Some publications, utilizing semi-empirical methods, suggest larger amounts of global mean sea-level rise by the end of the 21st century, reaching values in excess of 100 cm (e.g. 75 to 190 cm, Vermeer and Rahmstorf, 2009 and 57 to 110 cm, Jevrejeva et al., 2012). The semi-empirical projections are based on assumed relationships between global sea-level and either global temperatures or atmospheric heat balance. They do not capture the full range of physical processes responsible for changes in sea level. At present it is not known why they give higher values of sea-level rise than the process-based modeling which formed the basis of the IPCC TAR (Third Assessment Report) and AR4 (Fourth Assessment Report) results. It has been suggested that the semi-empirical projections be treated with caution owing to a number of limitations (Church et al., 2011).

An upper bound of 200 cm of global sea-level rise by 2100 was derived from glaciological modeling, to help rule out even larger values of global sea-level rise (Pfeffer et al., 2008). Based on an assessment of the probable maximum contributions from various sources of sea-level rise, and on studies using a variety of approaches, a "plausible high-end"

[5] The IPCC recently updated projections of future sea level rise (Stocker et al., 2013) and confirmed that higher levels of sea level rise (>1m) could not be excluded, but assessed the likely (>66% probability) upper end of the range of projected sea level rise to be about 1m relative to current levels by the end of the century.

scenario range of 55 to 115 cm of global sea-level rise by 2100 has been derived for use in flood risk planning (Katsman et al., 2011).

Included in these estimates are the contributions from melting Canadian glaciers and ice caps. The contribution to sea-level rise to 2100 from circumpolar Arctic glaciers is projected to be 5 to14 cm, with Canadian Arctic glaciers, ice fields, and ice caps projected to contribute 1 to 4 cm (AMAP, 2011). A recent update indicates a contribution of 3.5 ± 2.4 cm to global sea-level rise from the Canadian Arctic Archipelago in the 21st century (Lenaerts et al., 2013). Much smaller contributions are expected from western Canadian glaciers given their lower ice volumes (Marzeion et al., 2012).

The patterns of future relative sea-level change in Canada, like past and present-day patterns, will be influenced by land uplift and subsidence, uneven melt-water redistribution, and changes in ocean temperature, salinity and circulation (e.g. Slangen et al., 2012). Around Hudson Bay, some coastlines are rising so quickly that sea-level will continue to fall throughout the 21st century, except for the most extreme scenarios of global sea-level rise (James et al., 2011). A consequence of sea-level fall is reduced depth-under-keel of ocean-going vessels, leading to potential navigation and docking hazards. Areas that are rising more slowly may experience a transition from relative sea-level fall in the early decades of the 21st century to sea-level rise by 2100, depending on the rate of uplift and the amount of global sea-level rise. Subsiding regions will experience enhanced sea-level rise.

Melt-water redistribution in the oceans is uneven (Mitrovica et al., 2001, 2011). The shrinking mass of ice sheets and glaciers reduces their gravitational attraction to water in the oceans, leading to sea-level fall close to a source of meltwater. Near the source of meltwater, the Earth's crust responds elastically to the decreasing load, causing land uplift that also contributes to relative sea-level fall. These processes of meltwater redistribution and elastic crustal response (sometimes termed 'sea-level fingerprinting') are important in Canada because of the presence of Arctic ice caps and, in the west, mountain glaciers and ice fields. In addition, the Greenland ice sheet and Gulf of Alaska glaciers are both sources of meltwater for global sea-level rise. Due to their proximity – on a global scale – to these important sources of melt-water, large regions of Canada will experience reduced rates of relative sea-level rise. The effects of meltwater redistribution are sufficiently pronounced in parts of Arctic Canada that the range of local sea-level projections is less than half the range of global projections (James et al., 2011).

Sea levels are also affected by global ocean circulation, which accounts for greater than 2 m of current spatial variation in absolute sea level. The largest sea level gradient off the coast of Canada is located in the northwest Atlantic where the sea level change across the Gulf Stream is about 1.5m (Thompson et al., 2011). Variability in ocean currents may contribute to sea level change on all three coasts. Above-average sea-level risedue to changes in ocean circulation is projected in the Arctic and the Maritimes (e.g. Yin, 2012; Ezer et al., 2013), partly counteracting the reductions arising from melt-water redistribution. Off the west coast, long-term current-induced changes in coastal sea level may be masked by decadal-scale variations in sea level arising from changes in circulation and upper ocean temperatures associated with major El Niño and La Niña events (Thomson et al., 2008).

Projections of global sea-level rise beyond 2100 have an even larger uncertainty, but indicate continuing global sea-level rise over the coming centuries and millennia (e.g. Katsman et al., 2011; Huybrechts et al., 2011; Jevrejeva et al., 2012). Global sea-level rise may eventually amount to several metres.

5.4.3 EXTREME WATER LEVELS

Rising mean sea levels are an important factor with respect to extreme (high) water levels, which generally occur when storm surges coincide with high tidal levels. Contributions from harbour seiches, wind waves, and interannual and seasonal variability are also important. Ocean-surface heights vary on time scales from years to hours due to atmosphere and ocean variations, such as ENSO, NAO, seasonal warming and runoff, storms, and changes to ocean circulation. In the Pacific, extreme ENSO events can result in coastal sea level changes of a few tens of centimetres. Storm surges can have amplitudes of more than a metre on all three coasts (Bernier and Thompson, 2006; Manson and Solomon, 2007; Thomson et al., 2008). This short-term, large-amplitude variability causes peak water levels to vary substantially throughout the year and from year to year. It is superimposed on the slow rise in mean sea level which causes incrementally higher water levels over time where relative sea level is rising. In the Bay of Fundy, increasing mean sea level is resulting in a small increase in the tidal range due to increased resonance of the

semidiurnal tides (Greenberg et al., 2012), which will further contribute to extreme high water levels there.

Climate-related changes in the above factors will also affect extreme water levels in many regions of the globe. Possible climate changes affecting the intensity and frequency of storms, hurricanes and high wind waves are of particular concern, though they are expected to vary geographically and there is uncertainty regarding their sign and magnitude in most areas (e.g. Ulbrich et al., 2009; Harvey et al., 2012; Rummukainen, 2012; Seneviratne et al., 2012). Available analyses of observed wind speed changes at coastal locations around Canada are inconclusive regarding long-term trends (Hundecha et al., 2008; Wan et al., 2010). There are some suggestions that the strongest storms will become more intense in mid-to-high latitude areas of the North Pacific and North Atlantic (e.g. Mizuta, 2012; Woollings et al. 2012), associated with poleward shifts of the jet stream and storm tracks. However, there are differences in the details of the projected changes depending on season and location (e.g. Perrie et al. 2010; Long et al. 2009) and among models.

Changes in sea-ice cover have important implications for wind waves reaching the coast. Nearshore sea ice prevents waves from breaking directly onshore and reduces wave run-up (Forbes and Taylor, 1994; Allard et al., 1998). Ice further offshore reflects waves and reduces the amplitude of waves before they reach the shoreline (Wadhams et al., 1988; Squire, 2007), so that more open water will lead to larger waves even if the winds are unchanged. Thus, where there are projected reductions in sea ice, such as Atlantic Canada and the Arctic, there is the potential for increased extreme water levels due to run-up.

Increased extreme water levels will generally lead to increased amounts of coastal erosion. Dyked areas, coastal regions with little relief, and coastlines comprised of unconsolidated sediments are more vulnerable to erosion than high-lying, rocky coastlines. In the Arctic, increased air and water temperatures may degrade and thaw permafrost, loosening ice-bonded sediments and also contributing to erosion (Forbes, 2011). At this time, it appears that the long-term changes to the frequency and intensity of extreme coastal water levels and flooding in Canada will be primarily driven by changes in mean sea level and by sea ice changes, although tides, storm surge, and waves will continue to play prominent roles. Regions that are projected to experience an increase in mean sea-level are also likely to experience increasing extreme high water levels.

6. SUMMARY

Atmospheric warming has been widespread across Canada since 1950, although strongest in the north and west. It has occurred in all seasons, but has been most pronounced in winter and spring. The primary contributor to long-term warming in Canada (and the rest of the world) since the mid-20th century has been the anthropogenic emission of GHGs. Other factors can strongly influence short-term climate variability imposed on the long-term trend.

A range of indicators provides a coherent picture of the response of the atmosphere-ice-ocean system to this climate warming. An increase in hot extremes and a decrease in cold extremes of air temperature have been observed across the country. Canada as a whole has become wetter, although with notable spatial and seasonal variability. In most of southern Canada there has been a decrease in snowfall and an increase in rainfall consistent with warmer temperatures. A reduction in the spatial extent and mass of the Canadian cryosphere is evident in observations of rapidly declining snow and sea ice cover, shorter seasons of ice cover on many lakes and rivers, widespread warming of permafrost and shrinkage of glaciers in both western Canada and the High Arctic. Indicators of surface freshwater availability, such as streamflow, provide integrated responses to climate and cryospheric change, but spatially consistent patterns across the country are difficult to discern.

Natural climate fluctuations such as El Niño and the North Atlantic Oscillation contribute to regional climate variability on short (decadal) time scales. The warming projected to occur throughout this century will be associated with a continuation and potential acceleration of many of the trends observed over the past half century. Some patterns of change may prevail for Canada as a whole (a warmer, wetter Canada with less snow and ice), but regional and seasonal variability will continue. In particular, amplified warming and related impacts in the Arctic are expected. Precipitation changes are particularly uncertain, but potential declines in southern Canada, combined with warmer summers and increased evaporation, could increase seasonal aridity and reduce freshwater availability in some areas.

Long-term changes in ocean climate – temperature, salinity, oxygen levels and acidity – consistent with increasing atmospheric CO_2 and anthropogenic climate warming have been observed in all three of Canada's oceans. However, natural variability on decadal to multi-decadal time scales has also contributed to the observed changes in some areas (e.g. the Northwest Atlantic) off Canada. Nevertheless, warmer waters, reduced sea ice, reduced upper ocean salinities, and increased vertical density stratification are expected in most Canadian waters over the next century. The observed global trends of ocean acidification and reduced sub-surface oxygen levels are expected to continue and to be evident in Canadian waters as well.

Sea level change along Canadian coastlines has been, and will continue to be, affected by both global and local factors. Expansion of warming waters and increased meltwater from land ice are both contributing to rising global sea levels. Estimates of the magnitude of future changes in global sea level by the year 2100 range from a few tens of centimetres to more than a metre. Vertical land movement strongly influences relative sea level changes at the local scale. Where the land is currently subsiding, such as most of the Maritimes, relative sea level is rising at rates larger than the global average, and will continue to rise. Where the land is rising rapidly (e.g. around Hudson Bay), sea level will continue to fall except under extreme scenarios of sea level rise. Areas where land is rising more slowly may see a transition from relative sea level fall to relative sea level rise over the 21st century. Extreme sea levels are likely to be experienced more frequently in the coming century where relative sea level is rising and where sea ice is projected to decrease in the Arctic and in Atlantic Canada.

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