
Tracing Freshwater Anomalies Through the Air-Land-Ocean System: A Case Study from the Mackenzie River Basin and the Beaufort Gyre

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ABSTRACT Mackenzie River discharge was at a record low in water year (WY) 1995 (October 1994 to September 1995), was near average in WY 1996, and was at a record high in WY 1997. The record high discharge in WY 1997, with above average flow each month, was followed by a record high flow in May 1998, then a sharp decline. Through diagnosing these changing flows and their expression in the Beaufort Sea via synthesis of observations and model output, this study provides insight into the nature of the Arctic's freshwater system. The low discharge in WY 1995 manifests negative anomalies in P-E and precipitation, recycled summer precipitation, and dry surface conditions immediately prior to the water year. The complex hydrograph for WY 1996 reflects a combination of spring soil moisture recharge, buffering by rising lake levels, positive P-E anomalies in summer, and a massive release of water held in storage by Bennett Dam. The record high discharge in WY 1997 manifests the dual effects of reduced buffering by lakes and positive P-E anomalies for most of the year. With reduced buffering, only modest P-E the following spring led to a record discharge in May 1998. As simulated with a coupled ice-ocean model, the record low discharge in WY 1995 contributed to a negative freshwater anomaly on the Mackenzie shelf lasting throughout the winter of 1995/96. High discharge from July–October 1996 contributed approximately 20% to a positive freshwater anomaly forming in the Beaufort Sea in the autumn of that year. The remainder was associated with reduced autumn/winter ice growth, strong ice melt the previous summer, and positive P-E anomalies over the ocean itself. Starting in autumn 1997 and throughout 1998, the upper ocean became more saline owing to sea-ice growth.

RÉSUMÉ [Traduit par la rédaction] Le débit du fleuve Mackenzie a connu un minimum record durant l'année hydrologique (AH) 1995 (octobre 1994 à septembre 1995), il était près de la moyenne durant l'AH 1996 et il a connu un maximum record durant l'AH 1997. Le débit maximum record de l'AH 1997, avec des débits au-dessus de la moyenne à tous les mois, a été suivi d'un débit maximum record en mai 1998, puis d'une diminution prononcée. En diagnostiquant ces débits changeants et leurs effets dans la mer de Beaufort par le biais d'une synthèse d'observations et de sorties de modèle, cette étude donne un aperçu des caractéristiques du réseau d'eau douce de l'Arctique. Le faible débit de l'AH 1995 est le résultat d'anomalies négatives dans l'indice P-E et dans les précipitations de même que des précipitations estivales recyclées et des conditions de surface sèches immédiatement avant l'année hydrologique. L'hydrogramme complexe pour l'AH 1996 est dû à une combinaison de la reconstitution printanière de l'humidité du sol, du stockage d'eau par la hausse du niveau des lacs, d'anomalies positives de l'indice P-E en été et du relâchement d'une importante quantité d'eau qui était retenue par le barrage Bennett. Le débit maximum record de l'AH 1997 reflète le double effet d'un stockage moindre par les lacs et d'anomalies positives de l'indice P-E durant la majeure partie de l'année. En raison du stockage réduit, le faible indice P-E le printemps suivant a donné lieu à un débit record en mai 1998. Comme cela a été simulé à l'aide d'un modèle couplé glace-océan, le débit minimum record de l'AH 1995 a contribué à une

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anomalie négative d'eau douce sur le plateau du Mackenzie qui a perduré durant tout l'hiver 1995/1996. Un débit élevé de juillet à octobre 1996 a contribué dans une mesure d'environ 20 % à une anomalie positive d'eau douce qui est apparue dans la mer de Beaufort à l'automne de cette année. Le reste a été attribué à une croissance réduite de la glace en automne et en hiver, à une fonte intense l'été précédent et à des anomalies positives de l'indice P-E au-dessus de l'océan même. À partir de l'automne 1997 et durant tout 1998, la salinité de la couche supérieure de l'océan a augmenté à cause de la croissance de la glace de mer.

1 Introduction

The goal of the National Science Foundation Arctic System Study Freshwater Integration (FWI) is to understand the Arctic's freshwater system better. Research under the FWI study has ranged widely. For example, efforts to assess the mean pan-Arctic freshwater budget and projected changes throughout the twenty-first century (Serreze et al., 2006; Holland et al., 2007) have been complemented by focused studies of contemporary trends and variability in individual components, such as discharge from the major Eurasian rivers (Peterson et al., 2002), potential effects of freshwater export from the Arctic Ocean on the thermohaline circulation (Rennermalm et al., 2006), as well as regional process studies. A lesson learned from the FWI study is the value of "following the water" through the atmospheric, land surface and ocean branches of the system by synthesizing information from observations and models. In this spirit, this study describes the fascinating story of how a series of net precipitation (P-E) anomalies over the Mackenzie watershed, translated through seasonal snowpack storage, runoff, artificial impoundments and the changing buffering capacities of Lake Athabasca and Great Slave Lake, manifested themselves as anomalous discharge at the mouth of the river, to meet their ultimate fate in the Beaufort Sea. Our paper complements studies conducted under the Mackenzie component of the Global Energy and Water Cycle Experiment (GEWEX). A special 2002 issue of *Atmosphere-Ocean* (vol. 40) contains many of the key results from the Mackenzie GEWEX Study (MAGS).

Terrestrial runoff to the Beaufort Sea is dominated by the Mackenzie River. With a length greater than 4200 km, the Mackenzie is the longest river in Canada, draining an area of some 1.8 million km² (Woo et al., 2008) and contributes approximately 7% of the annual inflow to the Arctic Ocean as a whole (Fig. 1). Based on the gauging station at Arctic Red River, near the mouth of the Mackenzie, annual mean runoff (R) for the period 1972–97 is estimated at 171 mm yr⁻¹. Runoff represents discharge divided by the 1.8 million km² upstream area. This 171 mm yr⁻¹ mean runoff compares to an annual mean precipitation (P, based on the available gauge network and accounting for gauge biases) of 411 mm yr⁻¹. The difference between P and R yields an annual evapotranspiration (ET) of 241 mm yr⁻¹ (Serreze et al., 2003a). Long-term annual mean P-E should approximately equal the vertically integrated convergence of water vapour over the watershed. Data from the National Centers for Environmental Prediction/National Center for Atmospheric Research (hereafter NCEP) reanalysis (Kalnay et al., 1996) have been shown

to provide useful information on seasonal and interannual variability in P-E over the Mackenzie and other large Arctic watersheds (Liu et al., 2002; Serreze et al., 2003).

Precipitation over the Mackenzie basin tends to have a cold season minimum and warm season maximum. Moisture associated with winter precipitation tends to be transported in from the southwest by extratropical cyclones. The effects of lee cyclogenesis are prominent (Lackmann and Gyakum, 1996; Misra et al., 2000). Bjornsson et al. (1995) find that above-average winter precipitation over the watershed is associated with a positive 700 hPa height anomaly in the Pacific and a negative height anomaly over northeast Canada, providing favourable conditions for cyclone development by increasing the lateral temperature gradient over the Mackenzie. Using 500 hPa composite fields, Serreze et al. (2003) found similar relationships with respect to basin-averaged winter P-E. The pattern associated with low precipitation and P-E (negative tropospheric height anomalies in the Pacific and positive anomalies over northeast Canada) projects strongly on the typical longwave flow associated with the warm phase of the El Niño Southern Oscillation (ENSO), similar to that associated with the positive phase of the Pacific North American (PNA) teleconnection pattern. The negative precipitation anomalies tend to be associated with positive anomalies in surface air temperature.

The summer moisture flow into the basin tends to come from the northwest, the source region being the Arctic Ocean (Smirnov and Moore, 2001). While cyclone activity is still an important precipitation process, convective precipitation is also frequent. Despite the summer precipitation maximum, the high attendant evaporation rates result in a summer minimum in P-E, a feature shared by the major Eurasian watersheds (Walsh et al., 1996; Serreze et al., 2003a). In other words, relatively little of the summer precipitation tends to be available for runoff. Summer P-E can nevertheless be strongly positive for anomalous months. Szeto (2002) examined the precipitation recycling ratio in the Mackenzie basin. This refers to the ratio of precipitation derived from local evaporation to the total precipitation in the basin. The annual recycling ratio averaged over the basin was found to be about 0.25, highest from May–July (>0.4) and lowest in winter (<0.1). When the basin is under the influence of a large-scale low pressure system, much of the summer recycled precipitation falls in the northwestern parts of the basin where runoff ratios (R/P) are fairly high and thus enhance summer discharge. When the basin is under the influence of persistent high pressure, much of the recycled precipitation falls over

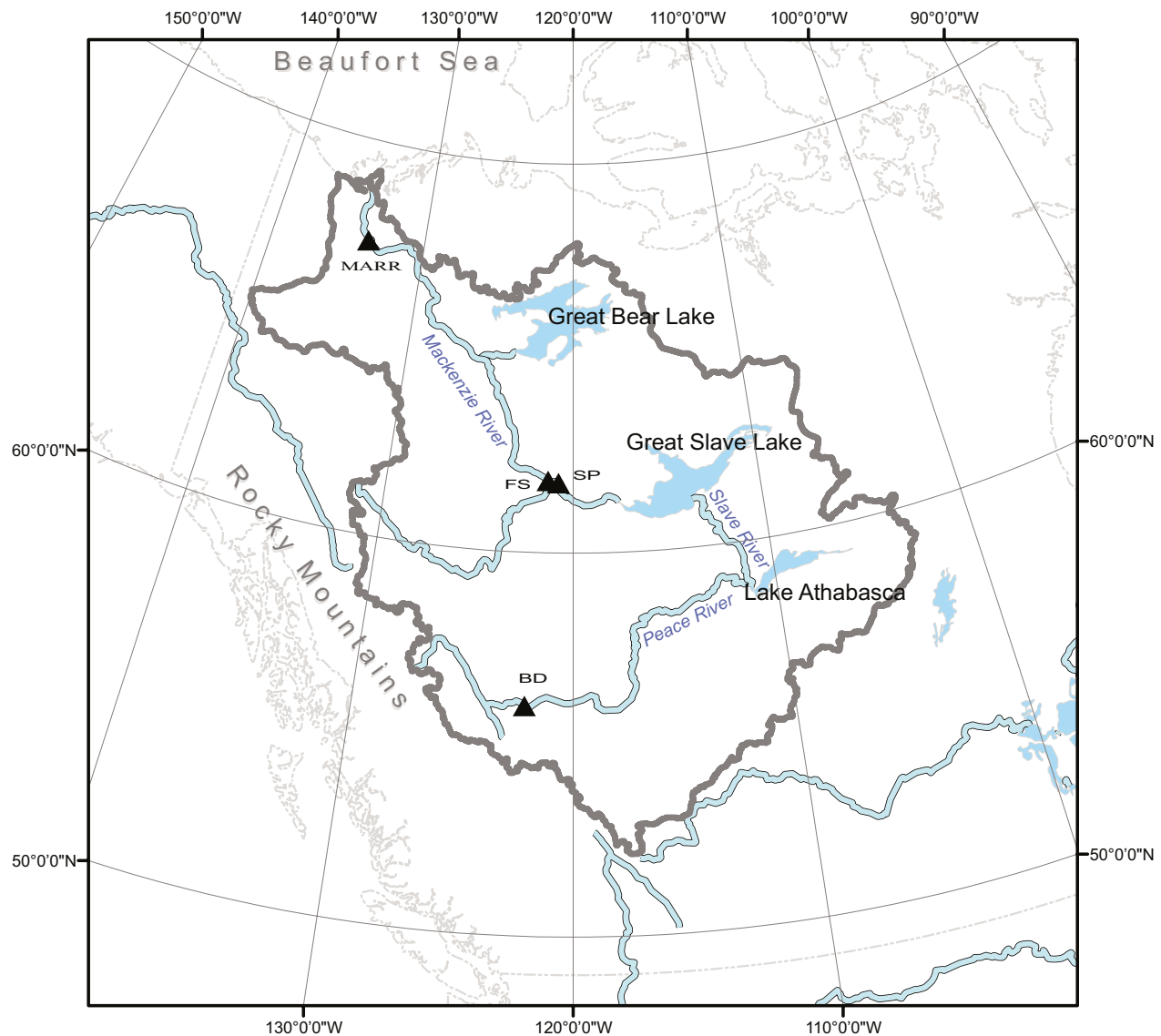


Fig. 1 The Mackenzie River Basin in northwest Canada (thick grey line). Triangles mark the locations of the discharge monitoring sites at Arctic Red River (MARR), Fort Simpson (FS), Strong Point (SP), and the Bennett Dam (BD).

the southern part of the basin where runoff ratios are lower, reducing discharge. Szeto (2002) suggests that this latter effect may have contributed to the record low summer discharge during 1995 described shortly.

The Mackenzie is similar to the major Eurasian Arctic drainages (the Ob, Yenisey and Lena) in that the bulk of runoff to the Arctic Ocean occurs from May–July, typically with a June peak. This is associated with melting of the winter snowpack; this snowpack can be viewed as a seasonal storage of P–E. However, the Mackenzie is unique among large Arctic-draining rivers in that it contains several large lakes along its course (Lake Athabasca, Great Slave Lake, Great Bear Lake) which can serve to modulate the flow of freshwater, sediment, and nutrients to the Arctic Ocean. The seasonality of flows is also modulated by dam operations, notably Bennett Dam, which impounds Williston Lake in the headwaters of the Peace River (Leconte et al., 2001).

Seasonality in river discharge is readily seen as low sea surface salinities in summer extending from the mouth of the river and along the Alaskan continental shelf. Surface salinities then tend to recover during autumn and winter as discharge declines, and sea-ice formation rejects brine into the underlying ocean. In turn, fresh water in both liquid form and as sea ice, tends to be transported away from the coastal region and into the central Arctic Ocean via the mean anticyclonic Beaufort Gyre. Assessing the fate of anomalous freshwater inputs from the Mackenzie is hence a complex problem, requiring consideration of the strong seasonality in the river input, regional atmospheric and oceanic conditions that effect the rate of ice formation and melt, as well as variability in the large-scale ocean circulation.

Following a description of the data sources and models, our study starts by examining characteristics of the atmospheric

circulation during the period 1995–98, links with anomalies in P–E over the Mackenzie basin, and variations in terrestrial snowpack storage. We then evaluate how these P–E anomalies, in conjunction with the changing buffering capacities of Lake Athabasca and Great Slave Lake, and water release from Bennett Dam, were translated into anomalous river discharge. Finally, we examine the fate of these anomalies in the Beaufort Sea. All anomalies are expressed with reference to the period 1980–2000. Annual discharge is based on the water year, defined as October until the end of September of the following year. For example, the 1995 water year extends from October 1994 until the end of September 1995.

2 Data and Methods

The NCEP reanalysis provides long time series of gridded atmospheric and surface fields. The most reliable fields such as pressure heights, tropospheric temperature and specific humidity are obtained by merging short-term atmospheric forecasts with observations. Terms of the surface energy budget such as downward longwave radiation and ground heat flux are less reliable as they are not directly influenced by observations of that variable. Temporal inconsistencies may arise due to changes in observing networks. NCEP fields are available from 1948, but since 1979, corresponding to the advent of modern satellite data streams, the fields are of higher quality. Use is made of fields of 500 hPa pressure heights, omega (vertical motion) at 500 hPa, and net precipitation (P–E), the latter calculated via the aerological method, adjusting the vertically integrated vapour flux convergence by the tendency in precipitable water, also known as column water vapour (Serreze et al., 2003b). Comparisons between different reanalyses for the polar cap suggest that it is reasonable to expect errors in aerological P–E of approximately 10% (Serreze et al., 2006). Across the Mackenzie basin the error may be larger given the region's complex topography. In our discussion, E represents the total loss of water from the surface such as open-water evaporation, transpiration from vegetation, and sublimation from snow. The P–E data used here represent the Mackenzie subset of 25×25 km gridded monthly products assembled as part of Arctic-Regional Integrated Hydrological Monitoring System (RIMS; <http://rims.unh.edu>).

Seasonal snowpack storage represents a major uncertainty. Direct measurements of snow water equivalent (SWE) are extremely sparse across Arctic lands and estimates derived from remotely sensed data are of generally poor quality. For characterizing SWE variability, simulations from land surface models (LSMs) forced with climate data (precipitation and air temperature) can provide a reasonable alternative to SWE fields derived from remote sensing (Rawlins et al., 2007). We use monthly estimates of SWE drawn from simulations with five LSMs (100 km grid array) forced with climate data from the European Centre for Medium-range Weather Forecasts Reanalysis (ERA-40; ECMWF, 2002) and applied across the pan-Arctic drainage basin. The five models generally produce a similar root mean square error, with median values below

40 mm. Biases among the models, however, are different, with two models having a median bias under 5 mm. Slater et al. (2007) provide additional details of the simulations. Here we use fields from the five-model SWE average.

River discharge records for stations along the Mackenzie are available from the R-ArcticNet v4.0 archive (www.r-arcticnet.sr.unh.edu/v4.0/). Records for most stations extend back to the early 1970s. We focus on the most downstream gauge, near the Mackenzie's confluence with the Arctic Red River (hereafter MARR). Record keeping began in 1973 (first complete year) and the time series has only five missing months for the period 1973–2000. For the missing months, December 1997 to March 1998, we make use of the linear regression between discharge recorded at MARR and discharge past the gauge at Fort Simpson, which are strongly correlated ($r \sim 0.8$). The missing November 1997 value was reconstructed by selecting analogous years with similar October discharge values at MARR. Uncertainties in annual discharge for the Mackenzie are likely similar to those from the large Russian rivers (Ob, Yenesei, Lena) which are less than 6% for annual discharge, approximately 20% for winter discharge, and near 10% during spring months when ice is present (Shiklomanov et al., 2006). Historical height data for the major Mackenzie basin lakes (used to assess their buffering capacities) are available from Environment Canada (<http://scitech.pyr.ec.gc.ca/waterweb/main.asp>). Height data for Great Slave Lake and Lake Athabasca have been recorded since 1959 and 1956, respectively. We are not aware of any error estimates for these data.

The fate of discharge anomalies in the Beaufort Sea is assessed via an ice-ocean model. This component of our study both builds upon and is motivated by the effort of Steele et al. (2006; hereafter SPZ06). This paper examined the origins of a negative sea surface salinity anomaly observed in the southern Beaufort Gyre in the autumn of 1997, at the start of the Surface Heat and Energy Balance of the Arctic (SHEBA) experiment, but using monthly climatological, rather than observed discharge from the Mackenzie (as well as from other rivers). SPZ06 found that the major contributors to the anomaly were enhanced sea-ice melt and Mackenzie River water. With regard to the latter, they found that the usual salinification of riverine shelf water by ice growth was severely curtailed in the autumn of 1996 by anomalous convergence of sea ice into the area from wind forcing, creating an anomalously thick, high concentration ice pack that limited ice growth and brine rejection into the ocean. The result was that the riverine contribution was retained largely as liquid over the winter of 1996/97, and was then advected northwestward in the summer and autumn of 1997 into the SHEBA deployment location.

River input was actually quite variable for 1995–98, with a record low in WY 1995 (126 mm yr^{-1}), average flow in WY 1996 (163 mm yr^{-1}), and a record high for WY 1997 (205 mm yr^{-1}) (Fig. 2) followed in turn by a record high in May of 1998 (see later discussion in Section 3a). This prompted us to develop two new simulations. The first is a

Water Year P–E and River Discharge Over the Mackenzie Basin

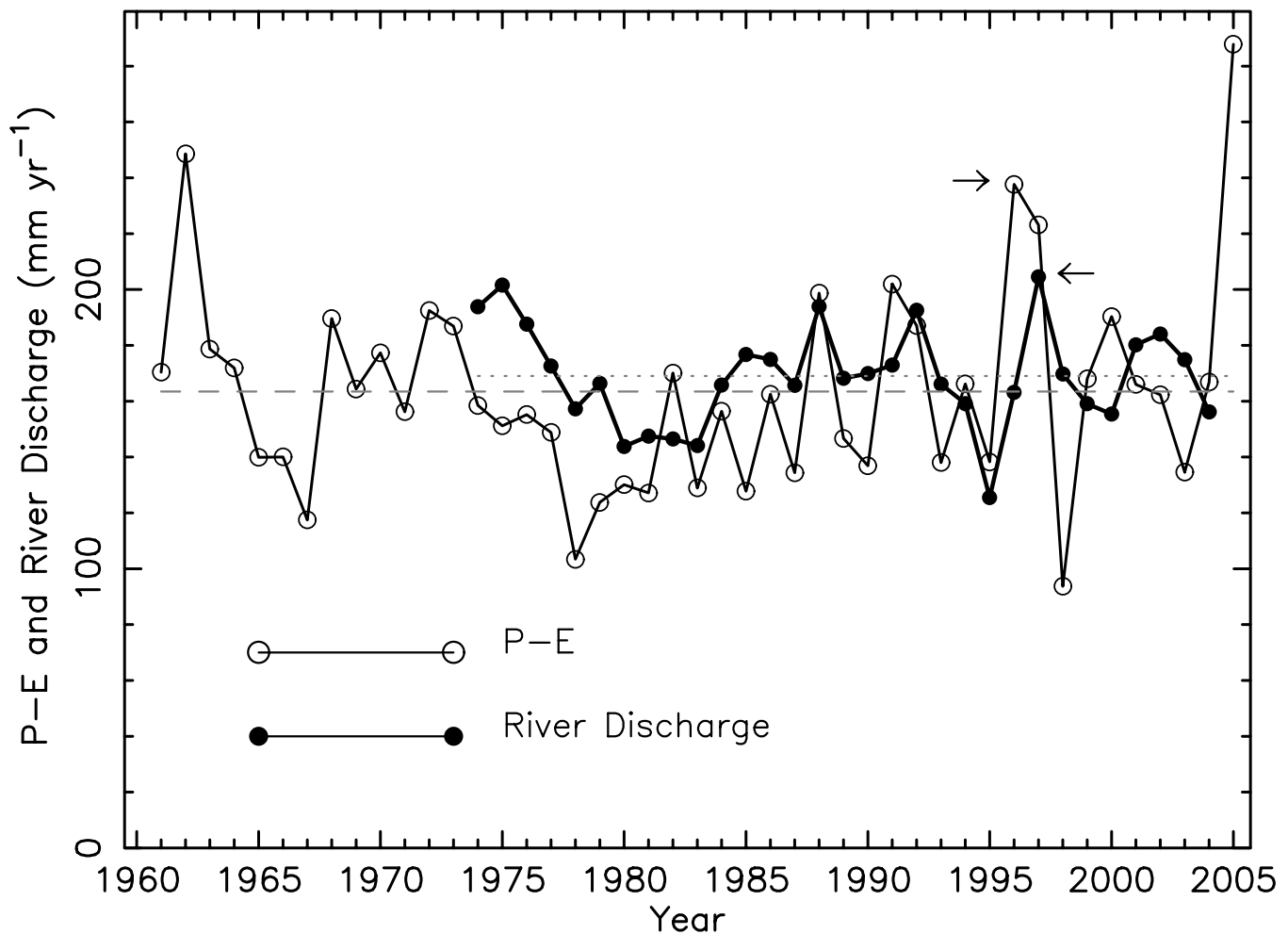


Fig. 2 Total discharge (1974–2004) observed at Arctic Red River and total P–E (1961–2005) averaged over the Mackenzie basin, based on the aerological method using data from NCEP. Plotted values represent water year (October–September) totals. Dashed and dotted lines show mean P–E and mean discharge over the respective periods. P–E for WY 1995/96 and river discharge for WY 1996/97 are indicated with right and left arrows, respectively.

control simulation in which each monthly discharge is simply the climatological monthly mean for MARR for the period 1973–2004. The second variable simulation uses the observed discharge time series. In both simulations, discharge for all other major Arctic rivers is represented by climatology, using values cited in Hibler and Bryan (1987). Climate forcings are taken from daily NCEP fields and the model is run for the period 1978–2004. This enables us to focus exclusively on the effects of Mackenzie River discharge variability on the Beaufort Gyre.

The ice-ocean model has also been further improved since the work of SPZ06. In particular, the ocean vertical mixing scheme now uses the K-profile parametrization (KPP) scheme (Large et al., 1994) in which mixing (diffusion and viscosity) is enhanced in the surface boundary layer and then reduced to a background value in the ocean interior. Zhang

and Steele (2007) discuss how this scheme improves simulation of the basic hydrographic structure in the Arctic Ocean, especially in the Beaufort Gyre region and also how it eliminates the need for any climate restoring (SPZ06 used constant mixing values throughout the water column and a climate restoring on salinity below 200 m). The model also has an enhanced vertical resolution of 30 z-levels, with 5 m resolution in the uppermost six levels (as opposed to the 10 m resolution in SPZ06).

Prior to analysis, the monthly and water-year discharge and P–E, and the monthly model SWE, were assessed for temporal autocorrelation using the Durbin-Watson (DW) statistic (Durbin and Watson, 1950) and by examining residual plots. For water-year discharge, the test was inconclusive. The DW values of the remaining time series are above the upper bound of the rejection region, and the null hypothesis of no temporal

autocorrelation is accepted. In the analysis to follow we consider the samples to be independent and use the z-score as our test statistic. Anomalies are noted provided that z falls outside the critical region of a two-tailed test at the 90% confidence level (10% significance level).

3 Results

a *P–E and Discharge for 1995–98*

Figure 3a plots monthly discharge ($\text{km}^3 \text{ mo}^{-1}$) at the MARR gauging site for the years 1994–98. As is clear in this figure but more so in Fig. 3b which plots anomalies, the annual values shown earlier are associated with large variability from month to month. The record low annual flow for WY 1995 was largely the result of negative anomalies from June to September, with June and July standing out prominently. The near-average discharge in WY 1996 reflects strong compensation between small or negative anomalies from October to June and positive anomalies for July to September. In sharp contrast, positive discharge anomalies characterize every month of WY 1997, especially June to September. Following above-average discharge in October 1997, flows to March of 1998 were near normal, followed by a record high for the month of May 1998, then a return to negative anomalies for the remainder of the water year. A test of the monthly anomalies for significance level suggests that only the most extreme departures, such as this May record flow, are significant at the 90% level. Relatively high variability is present at the monthly time scale. For the total water-year discharge, anomalies for 1995 (record low) and 1997 (record high) are also significant at the 90% level.

As discussed in several papers in the Atmosphere-Ocean special volume devoted to MAGS, a number of factors contributed to the record low discharge of WY 1995. Using NCEP data, Liu et al. (2002) found that the vapour flux convergence (P–E) over the Mackenzie for WY 1995 was the lowest of the ten-year period 1987–96, associated with strong negative precipitation anomalies in spring and summer (Louie et al., 2002; Stewart et al., 2002). Another factor was the low efficiency at which vapour flux convergence was converted into discharge, in turn related to the pattern of summer precipitation recycling noted earlier, very warm summer conditions, and dry surface conditions that developed just prior to the water year (Stewart et al., 2002; Szeto, 2002).

However, looking at Fig. 2, the water-year total P–E of 138 mm for WY 1995 based on our analysis, while certainly below average, does not stand out as particularly low in the NCEP record. Over the 1987–96 period examined by Liu et al. (2002), our P–E estimates for WYs 1987 and 1990 fall slightly below this 1995 value. Figure 4 shows monthly P–E averaged over the basin for 1994–98 calculated from the NCEP data, along with anomalies with respect to the period 1980–2000. WY 1995 was associated with a mix of positive and negative anomalies in P–E, with prominent positive anomalies for March and August contrasting with strong negative anomalies for January, May and September. Each of the latter can be associated with an atmospheric pattern charac-

terized by a strong 500 hPa ridge over the Mackenzie River basin (not shown).

The unusual hydrography for WY 1996 also appears to reflect a suite of processes. Figure 4b shows that the first half of the water year (snowpack growth) was associated with positive P–E anomalies. This was, of course, reflected in the winter snowpack. Figure 5b shows the late winter (January–March average) SWE storage for 1996, as depicted by the suite of land-surface models. SWE anomalies in the southern part of the basin were nearly double the long-term mean, but were partly compensated for by below-average values in the north. More than 40% of the grid cells have an anomaly that is positive and significant at the 90% level. For the basin as a whole, the snowpack averaged over January–March was 11% above average. Note the contrast with the corresponding SWE anomalies for 1995 (Fig. 5a), which are negative nearly everywhere except for a small area in the north-central part of the basin. However, only 3% of the grid cell anomalies are negative and significant in 1995.

Why then was the spring discharge in 1996 below normal? Part of the answer may lie with replenishment of soil moisture deficits that developed in 1995. Furthermore, as illustrated in Fig. 6a, there was a pronounced springtime rise in the levels of Lake Athabasca and Great Slave Lake from the low levels observed in 1995. By August 1996, storage in Lake Athabasca and Great Slave Lake had increased by approximately 29 km^3 (nearly 10% of mean annual discharge at MARR) from February 1995 levels, and discharge at MARR was correspondingly reduced, with the discharge anomaly remaining negative to June 1996. Calculation of storage changes assumed no variation in horizontal extent in either lake. The total magnitude of the May–September 1996 discharge anomaly at MARR was $+13 \text{ km}^3$ compared to $+33 \text{ km}^3$ directly upstream of Lake Athabasca. The full lakes then contributed to elevated flow over the next several months. From Fig. 6b, it is clear that discharge in the open water season (May–October) is highly correlated with the level of Great Slave Lake. The buffering of approximately 10% of mean annual Mackenzie River discharge is likely conservative, as many other lakes such as Lesser Slave Lake and Lake Claire exhibited similar behaviour during 1996. With regard to these buffering effects, it is important to note that the strongly positive winter SWE anomalies were in the southern headwaters (Fig. 5b), immediately upstream of the lakes. If the SWE anomaly had instead developed in the lower reaches of the basin, below the lakes, we would not have seen this buffering effect.

Lake discharge was certainly not the sole contributor to the above-average discharge at MARR for the remainder of WY 1996. As is evident in Figs 4b, 7a and 7b, P–E anomalies were strongly positive in July and August. Largely as a result of these two months, P–E in WY 1996 was the second highest on record for the period 1960–2004 (Fig. 2). As is the case with the river discharge anomaly in 1997, the annual P–E anomaly in 1996 is significant at the 90% level. Figure 8

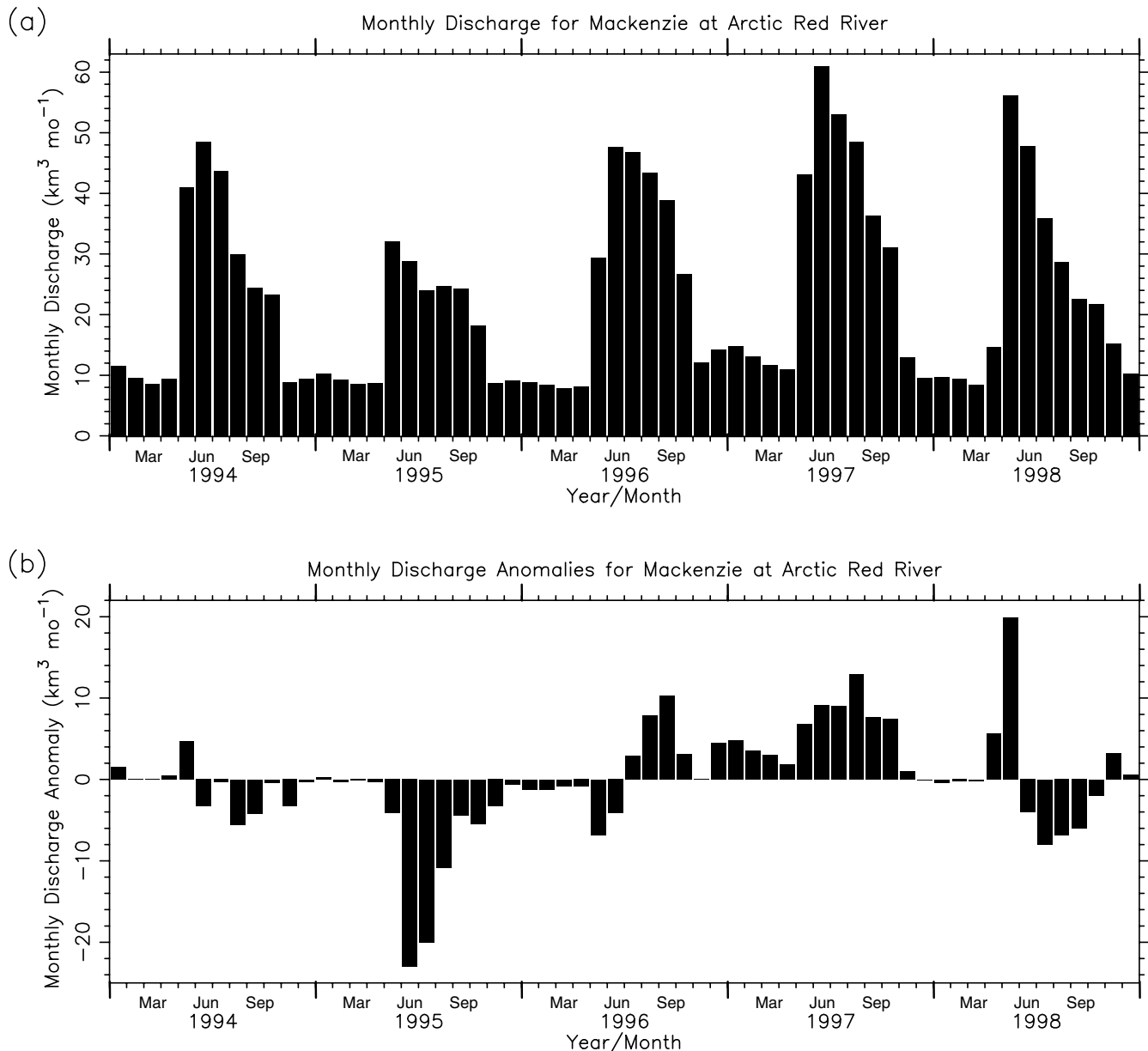


Fig. 3 Monthly discharge (a) and discharge anomalies (b) at Arctic Red River (MARR), 1994–98. Anomalies (here and in all subsequent figures) are computed with respect to the period 1980–2000.

shows the pattern of 500 hPa height anomalies associated with the unusual August 1996 P–E event. It is characterized by an anomalously strong ridge centred at the date line and a trough centred along the longitude of the Gulf of Alaska. Figure 8b depicts the anomaly pattern of omega (vertical motion) at 500 hPa. Negative omega means upward motion. While there is an obvious strong negative (upward) anomaly in the northern Gulf of Alaska in the expected position ahead of the trough where there will be positive relative vorticity advection increasing with height, anomalies of the same sign also extend into the Mackenzie basin. This vertical motion anomaly is consistent with the anomalously positive P–E.

Another contributor to the above-average discharge in 1996 was the release of a large quantity of water from Williston Lake impounded by the Bennett Dam. Over the period 14 June to 17 August 1996, there was an average release of about $5000 \text{ m}^3 \text{ s}^{-1}$, lowering the lake by about 3 m (Leconte et al., 2001). While contributing to high summer flows at Arctic Red River, the effects of this release can also be seen in the rising levels of Great Slave Lake and Lake Athabasca between July and August.

The impact of anomalously high river discharge in the summer of 1996 was particularly striking in the Peace-Athabasca Delta, a large wetland system on the western edge of Lake Athabasca. The ecology of the delta, which is

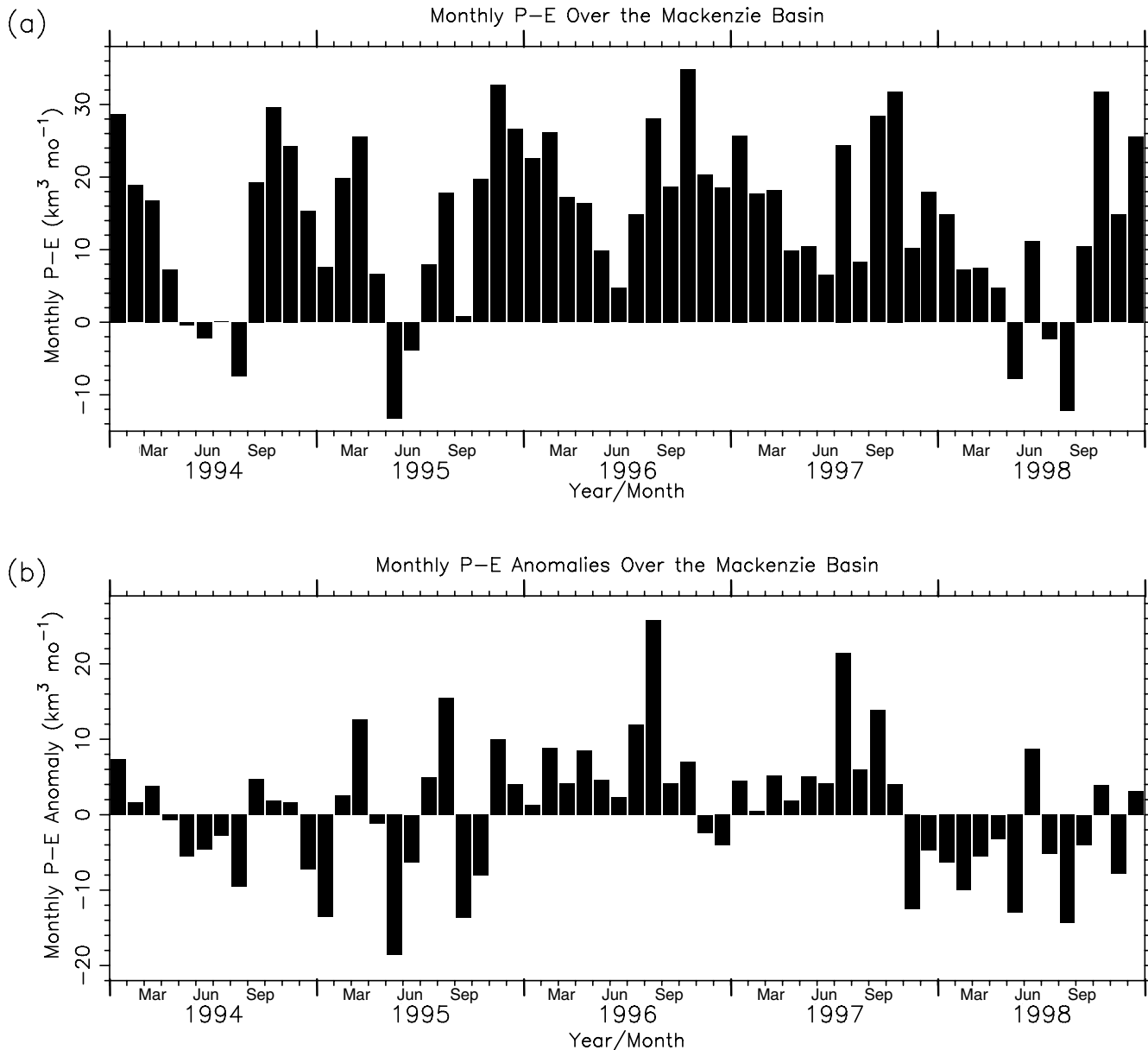


Fig. 4 Monthly (a) P-E and (b) P-E anomalies over the Mackenzie River basin for the years 1994–98.

recognized as a United Nations world heritage site because of its high biological diversity, is dependent on regular inundation through overbank flooding (Töyrä and Pietroniro, 2005). Prior to 1996, no significant flooding had occurred for a period of 22 years due to a combination of flow regulation and climatic variability (Prowse et al., 2006). Pervasive flooding beginning in May 1996 resulted in a recharge of hydrologically isolated lakes and wetlands that had not been inundated for more than two decades, restoring habitat for migratory birds and other species (Leconte et al., 2001).

Recall that discharge for WY 1997 was a record high. The first half of WY 1997 was associated with a mix of positive and negative P-E anomalies (Fig. 4b), expressed in January–March SWE as negative anomalies over most of the

basin, partly balanced by positive anomalies in the southern headwaters (Fig. 5c), a pattern broadly similar to that observed in 1996. Despite lowering lake levels during autumn and winter, lake levels were still quite high compared to normal in spring 1997 (Fig. 6a), and hence had limited ability to hold spring snowmelt. Therefore, despite the modest winter snowpack, there was a high spring discharge at the river mouth. Comparing Figs 2 and 4b, it is apparent that despite the mix of P-E anomalies for the first half of the water year, P-E for WY 1997 as a whole was only slightly lower than for WY 1996 (Fig. 2). The record high discharge can largely be explained by the very strong positive P-E anomalies in July, and to a lesser extent September, which helped to maintain high river flows during the summer and early autumn.

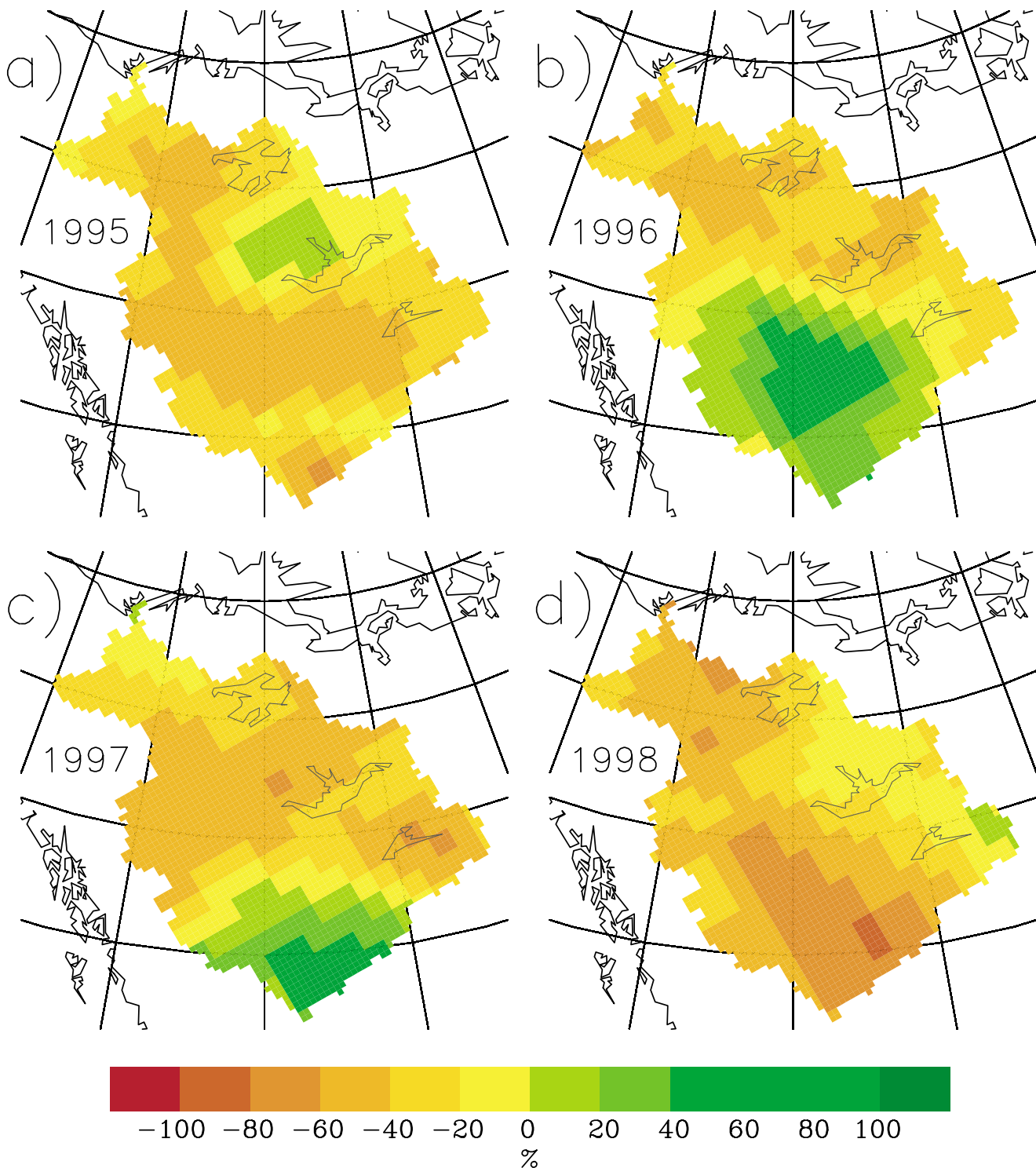
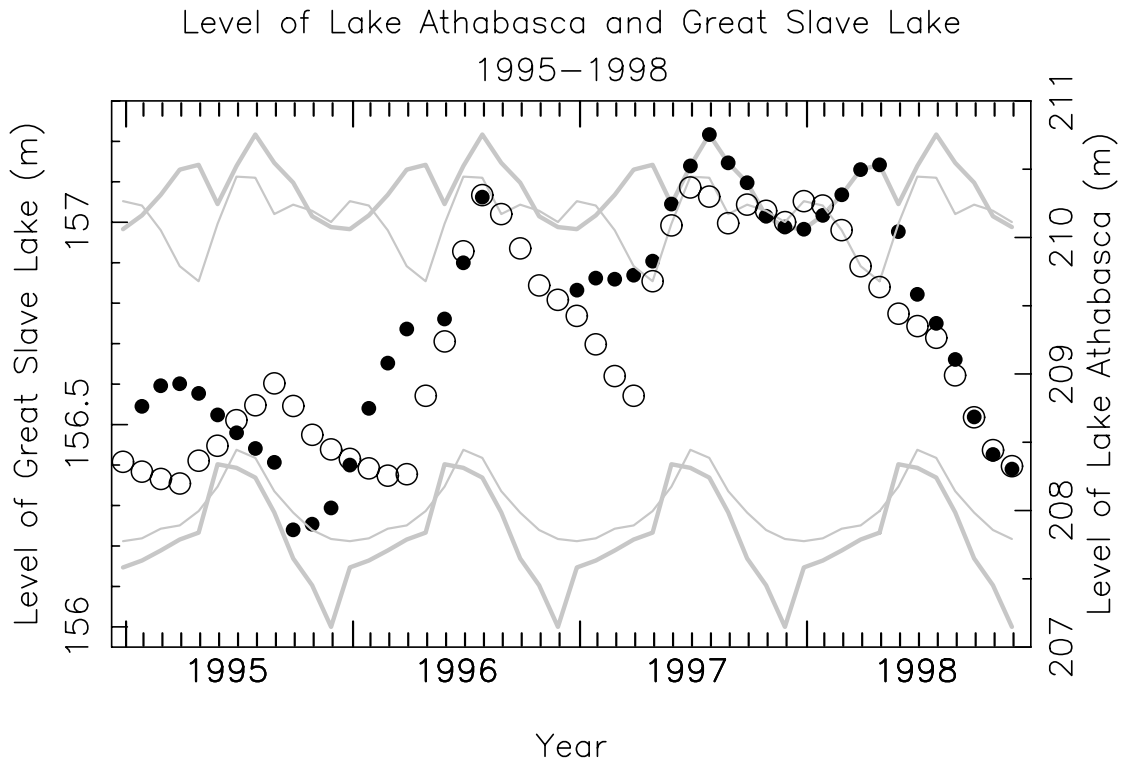


Fig. 5 Anomalies in January–March snow water equivalent across the Mackenzie basin based on averaging the output from five land surface models (see Slater et al. (2007) for models and setup) for (a) 1995, (b) 1996, (c) 1997, and (d) 1998.

This brings us to the interesting case of 1998. Referring again to Fig. 4, P–E anomalies remained negative from November 1997 to the spring of 1998. This is clearly an influence of the strong El-Niño event of 1997/98. Its expression is seen on the plots of 500 hPa height and anomalies

averaged for November to April (Fig. 9a). Note the strong negative height anomaly in the Gulf of Alaska and the positive anomaly over northern North America, extending into the Canadian Arctic Archipelago, associated with a pronounced western North American ridge. The pattern of positive

(a)



(b)

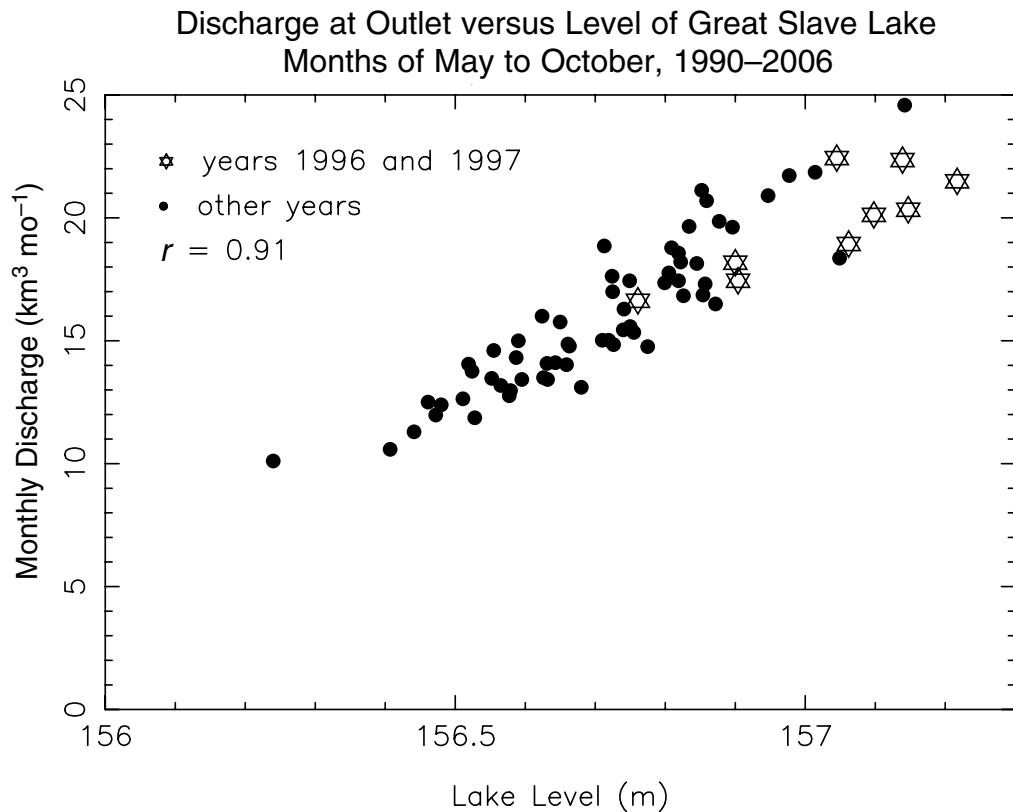


Fig. 6 (a) Levels of Great Slave Lake (filled circles, m) and Lake Athabasca (open circles) from Environment Canada data. Also shown are the minimum and maximum levels for Great Slave Lake (thick grey line) and Lake Athabasca (thin grey line). The historical minima and maxima shown represent monthly climatologies calculated over the period of record. (b) Relationship between lake level and discharge near the outlet of Great Slave Lake for the open season between May and October for the years 1990–2006.

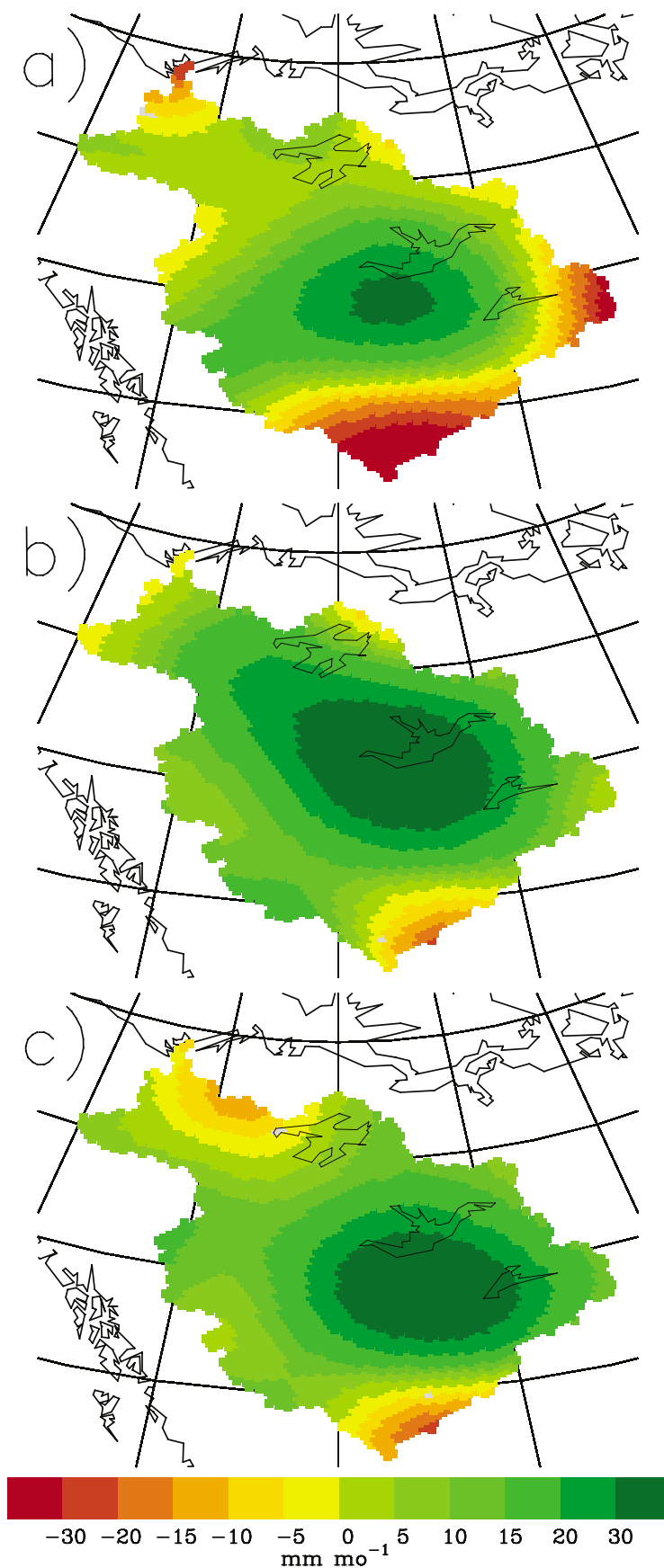


Fig. 7 Anomalies in monthly P-E across the Mackenzie basin based on NCEP data for (a) July 1996, (b) August 1996, and (c) July 1997.

Anomalies in 500 hPa Height and Omega

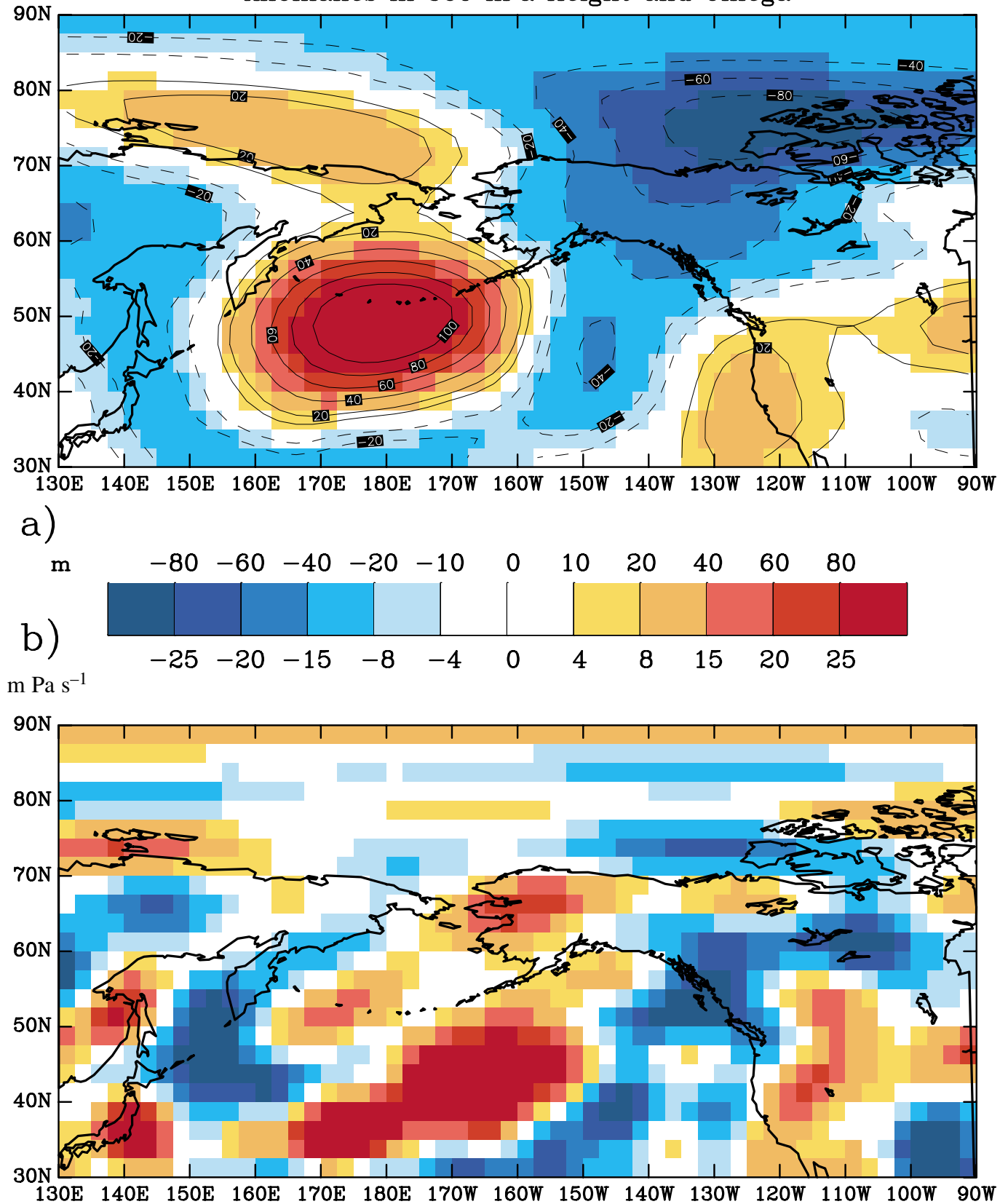


Fig. 8 Anomalies in (a) 500 hPa height and (b) omega (vertical motion) at 500 hPa associated with the unusually positive P-E in August 1996. Note the negative omega anomalies over the Mackenzie basin, associated with upward vertical motion.

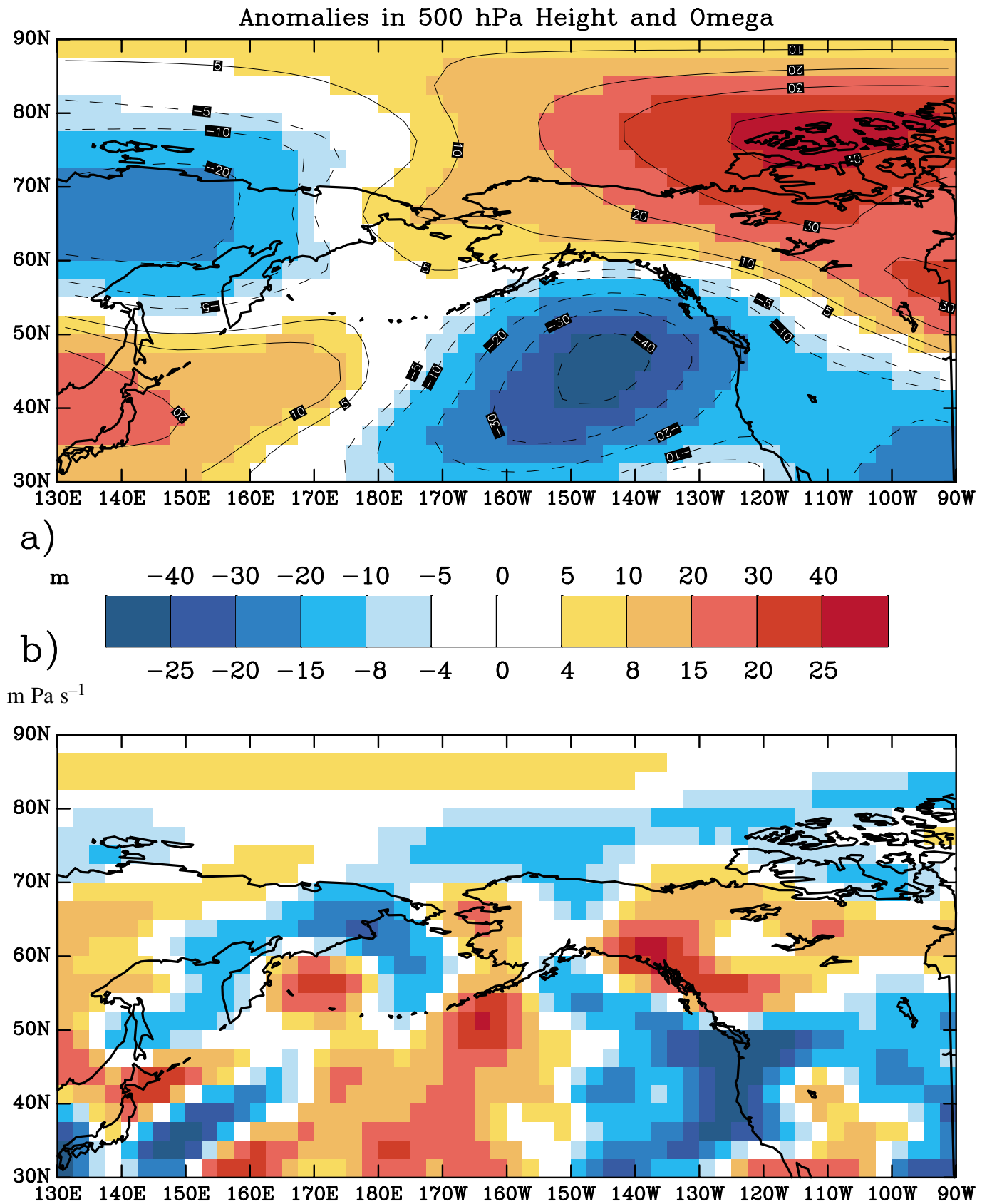


Fig. 9 Anomalies in (a) 500 hPa height and (b) omega (vertical motion) at 500 hPa averaged over the period November 1997 to April 1998. Note the positive omega anomalies over the Mackenzie basin, associated with downward vertical motion.

anomalies in 500 hPa omega over the Mackenzie basin for these months is revealing (Fig. 9b). Note the positive omega anomalies over most of the basin. Positive omega means downward motion, tending to suppress precipitation. These conditions are in turn consistent with below-normal winter SWE across almost the entire basin (Fig. 5d), expressed as a basin-averaged anomaly of approximately -22% .

Despite negative P–E and SWE anomalies, the May 1998 discharge at Arctic Red River was the highest on record for that month (approximately $56 \text{ km}^3 \text{ mo}^{-1}$). It seems that with the lake levels remaining very high in spring (Fig. 6a), hence with little buffering capacity, even below-normal P–E was expressed as very high May discharge. Discharge in subsequent months then fell below the 1996 and 1997 values in July to October, with the water-year value ending up as a modest $279 \text{ km}^3 \text{ yr}^{-1}$. In turn, the levels of Lake Athabasca and Great Slave Lake quickly retreated from their high spring values (Fig. 6a). These large variations are noteworthy given that climate change has contributed to an increase in the amplitude of water level variations, an increase in the annual water level maximum, and a modest advance in the timing of the water level maximum of Great Slave Lake (Gibson et al., 2006).

b *Dynamics of the Beaufort Sea Freshwater Anomaly*

How were these discharge anomalies at the mouth of the Mackenzie manifested in the Beaufort Sea? As a start to answering this question, Fig. 10 shows freshwater anomalies averaged over the upper 65 m from the simulation with inter-annually varying Mackenzie River discharge, displayed in a manner similar to that in Fig. 1 of SPZ06. Fresh water is defined as vertically integrated salinity, relative to the commonly used reference value of 34.8 taken to represent the average salinity of the Arctic Ocean (Serreze et al., 2006). The basic evolution of the anomaly is similar to that described in SPZ06. Discharge was below normal for WY 1995. Not surprisingly, this led to a negative freshwater anomaly on the Mackenzie shelf lasting throughout the winter of 1995/96. However, by the end of summer 1996 the anomaly was gone, freshened by anomalously strong summer ice melt. Now recall that water-year discharge was near average for WY 1996. Nevertheless, in the autumn of 1996, a positive freshwater anomaly formed along the Alaskan shelf and remained relatively stationary throughout the winter and early summer of 1997. It spread north and westward in the autumn of 1997 (Fig. 10a) and by autumn 1998 had been entrained into the anticyclonic Beaufort Gyre circulation (Fig. 10b). The anomaly remained in the Gyre for 3.5 years, attaining a peak amplitude of about 2 m in 1999 and 2000 (Figs 10c and 10d). The anomaly slowly drifted back towards the southern Beaufort Sea, as diffusion reduced its amplitude to below 0.5 m by 2002 (Figs 10e and 10f).

However, there are differences relative to SPZ06. First, the positive freshwater anomaly in the Beaufort Gyre in autumn 1997 (Fig. 10a) was absent in SPZ06. This Beaufort Gyre anomaly also exists in the control run, so it is not a result of

variable Mackenzie River discharge. Also, the anomaly lasts about one year longer in this simulation relative to SPZ06, probably a result of the improved mixing scheme discussed next.

The KPP mixing scheme used in the present simulation has improved the vertical structure of salinity (and salinity anomalies) relative to observations. This is apparent when comparing the salinity anomaly profiles from SPZ06 (their Fig. 3) with those of our interannually variable river discharge simulation (Fig. 11). The mixed layer structure is much more realistic in the present simulation. Figure 11 also shows that the freshwater anomaly during these years was confined to the upper 65 m. Also, while the surface salinity anomaly was essentially gone by autumn 2001, a subsurface anomaly is still evident in 2002. Comparison of the observed salinity anomaly profile at SHEBA in autumn 1998 with the nearby simulated profile at this time (red curve in Fig. 11) shows good agreement between roughly 20 m and 70 m depth. However, the modelled salinity is too fresh above approximately 15 m and below approximately 80 m depth. These differences are much reduced when averaging over the entire Canada Basin (Zhang and Steele, 2007).

What, then, was the effect of variable Mackenzie River discharge? This can be addressed by comparing the variable discharge run with the control run (with climatological river discharge). As in SPZ06, we focus on the autumn of 1996, when the low salinity anomaly formed on the Mackenzie shelf. Mackenzie River discharge was anomalously high from July to October 1996 by a total of approximately 19 km^3 (Fig. 3b). Thus we expect that the variable discharge run would produce more oceanic freshening relative to the control simulation. Figure 12a shows the upper 65 m freshwater anomaly from the variable discharge run just after it formed in the autumn of 1996. The maximum value on the Mackenzie shelf is about 1.6 m, with an average over the anomaly (black outlined area) of about 1.1 m. Figure 12b shows the difference in autumn freshening between the variable and control simulations. The variable discharge simulation indeed freshens more, about 0.18 m averaged within the black contour in the figure, with a maximum of 0.39 m. The net effect is an increased freshening of $0.18/1.1 = 16\%$ averaged over the Mackenzie shelf, or a value of $0.39/1.6 = 24\%$ using anomaly extremes. However, by the time the anomaly entered the Beaufort Gyre in 1999 and 2000, contributions from other sources (notably, sea-ice melt/growth anomalies) and dissipation reduced this value to about 2%.

SPZ06 found that existing heavy ice conditions during autumn 1996, leading to reduced ice growth and brine rejection, accounted for a freshwater anomaly of about 30 cm over $150,000 \text{ km}^2 = 45 \text{ km}^3$ (see their Fig. 4b). This compares with the previously mentioned river discharge anomaly of about 19 km^3 . Also, the local oceanic P–E anomaly in the southern Beaufort Gyre at this time was small, about 10% of the sea-ice forcing. Thus the river discharge forcing was about $19/(19+45+4.5) = 28\%$ of the total freshwater forcing. This is larger than the 16–24% ocean response depicted in Fig. 12b.

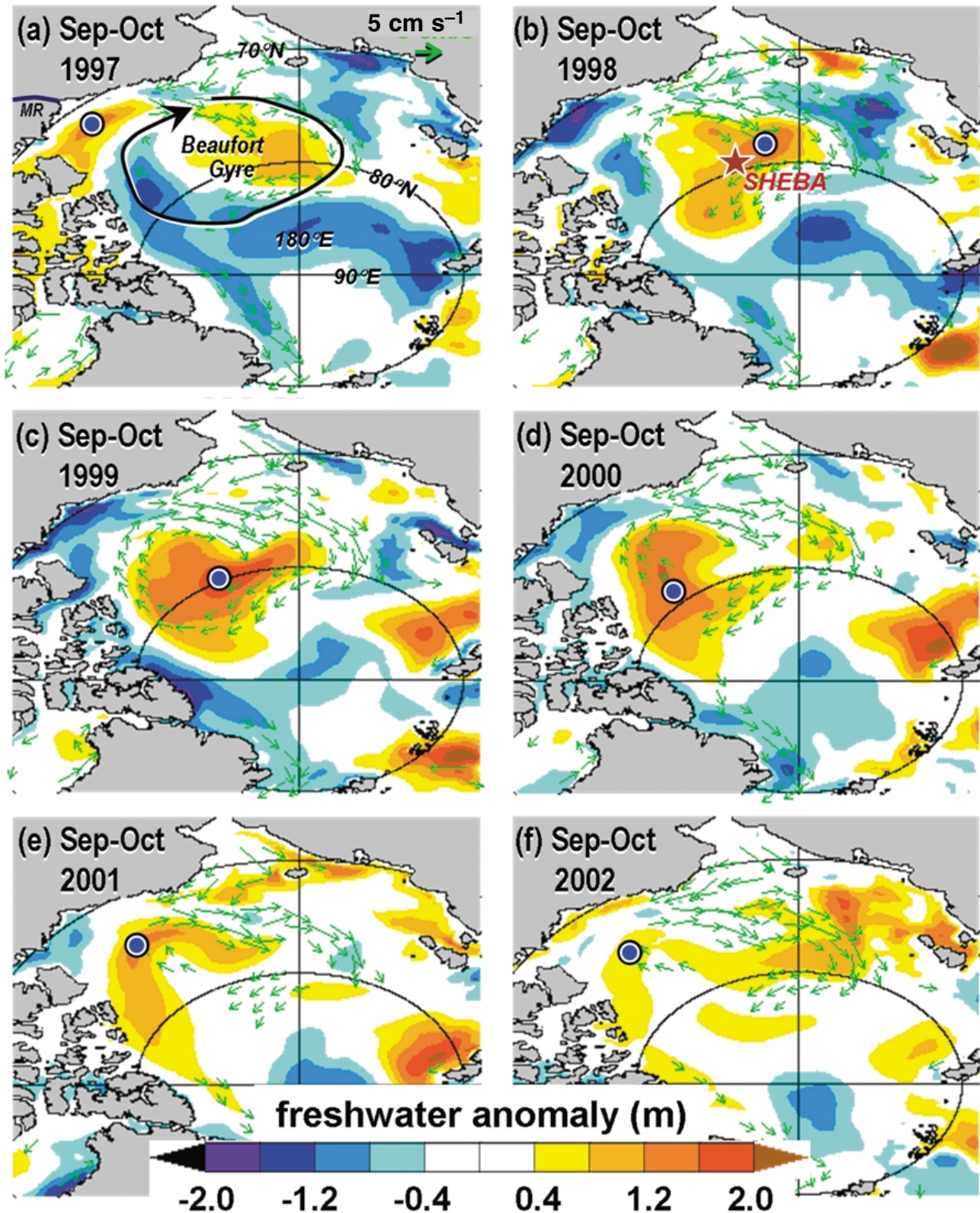


Fig. 10 Vertically integrated freshwater anomalies averaged over the upper 65 m (relative to 1980–2000 bimonthly means) during September–October of (a) 1997, (b) 1998, (c) 1999, (d) 2000, (e) 2001, and (f) 2002. Also shown are ocean currents averaged over this same depth, only plotted where they exceed 2 cm s^{-1} . A blue dot at the centre of the high freshwater anomaly marks the location of the profiles shown in Fig. 11. The dot in panel (f) is simply placed at the same location as in panel (e). Also marked are (a) the Mackenzie River (MR), and (b) the site of the SHEBA ice camp (star) at this time.

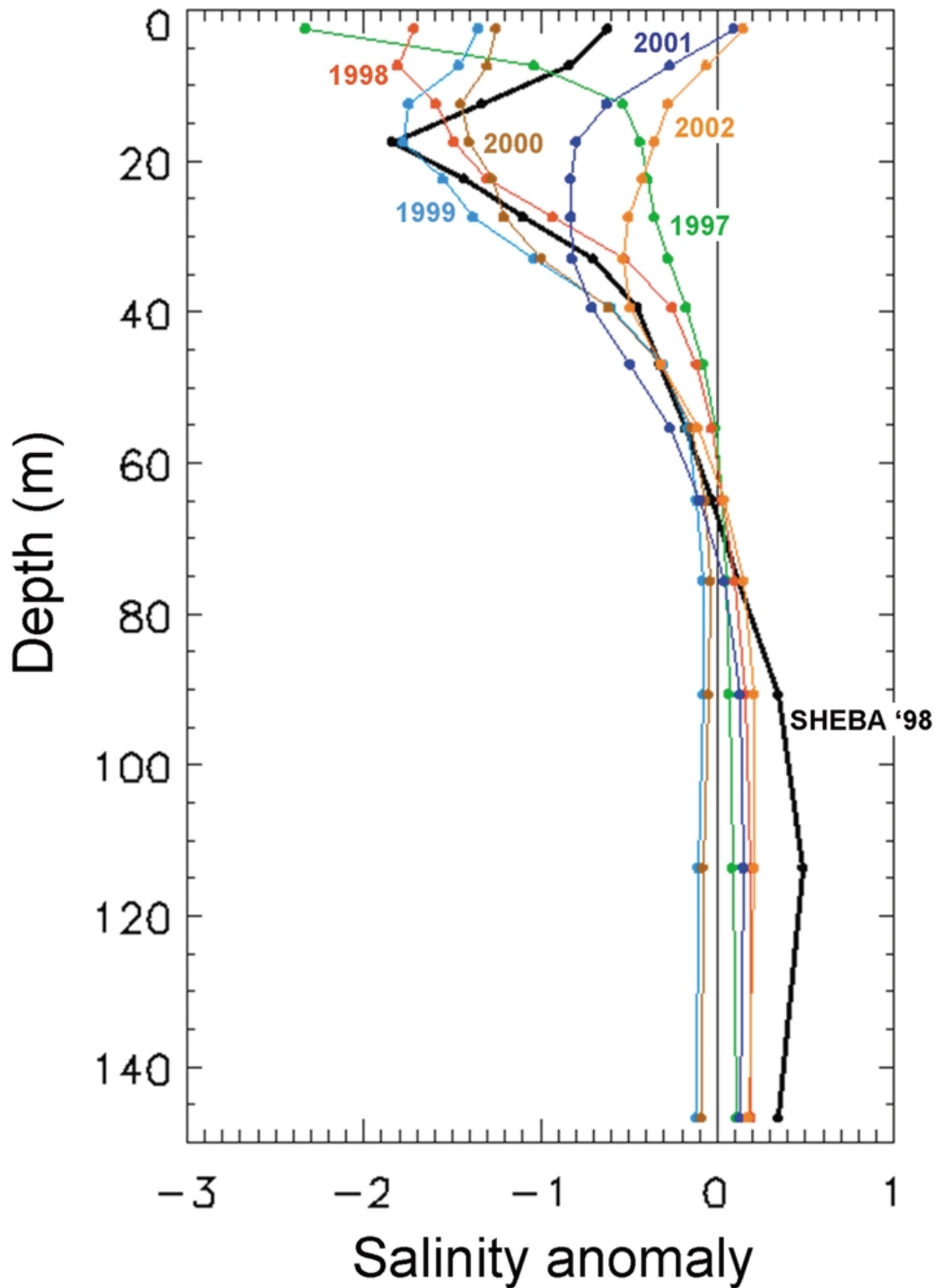


Fig. 11 Modelled mean September–October salinity anomaly profiles (relative to the 1980–2000 September–October mean) at the locations of the blue dots in Fig. 10. Also shown is the mean observed salinity profile at SHEBA during autumn 1998, minus the model mean at this location. Model depth levels appear as dots on each profile.

The model might be underestimating the freshwater anomaly owing to uncertainties in parameterizing horizontal and vertical mixing (Zhang and Steele, 2007; Golubeva and Platov, 2007) or our use of a common but simplified salinity boundary flux formulation in which mass flux is neglected (Tartinville et al., 2001).

Mackenzie River discharge was at a record high for WY 1997 and for the month of May in 1998 (Fig. 3b). However, discharge was near the climatological mean for much of the 1997/98 winter, when ocean freshwater anomalies on the Mackenzie shelf turned negative. They remained negative until 2002 (Figs 10b–10e), owing to anomalously low

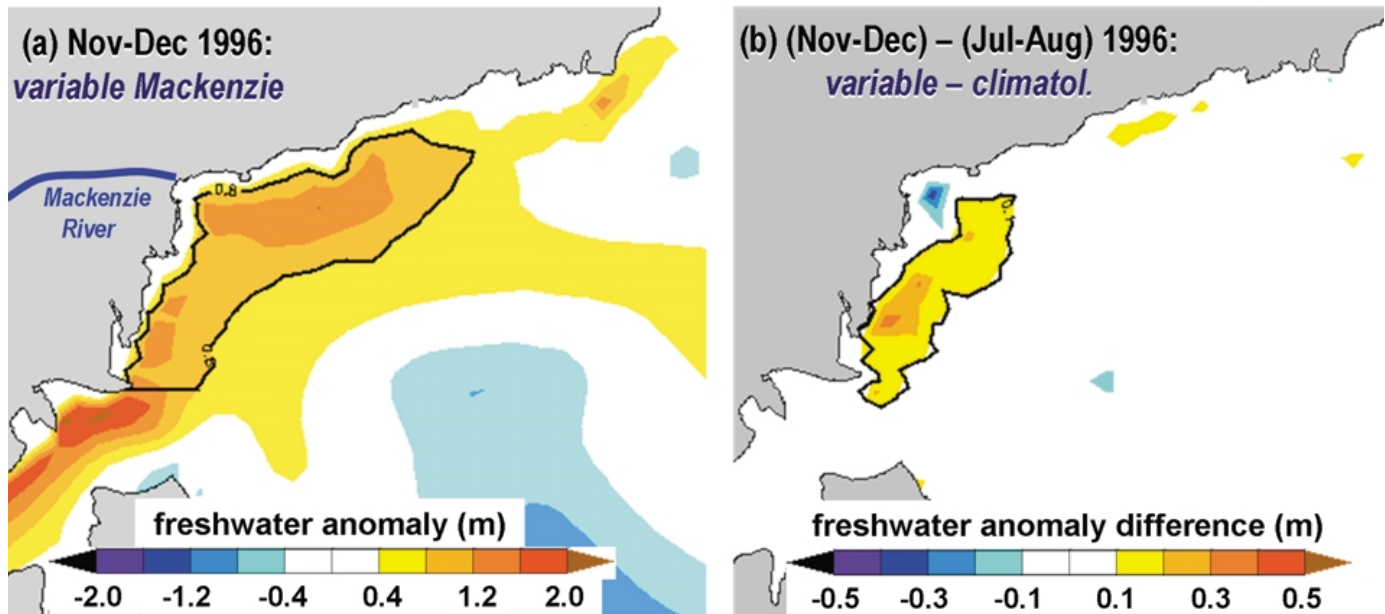


Fig. 12 (a) November–December 1996 upper 65 m freshwater anomaly (relative to the 1980–2000 mean) for the simulation with interannually varying Mackenzie River discharge. (b) The difference between autumn and summer freshwater change (i.e., November–December minus July–August) in the variable minus the climatological Mackenzie River discharge simulations.

summer sea-ice melt (sea ice has a very low salinity) and high autumn sea-ice growth (increased brine rejection), especially from summer 1997 to summer 1998. During this period, the region acted something like a polynya, with net ice growth balanced by net divergence of sea ice away from the region towards the northwest (see SPZ06 Figs 2 and 5). Thus, it seems that the years 1996 and 1997 offer an interesting contrast in how river discharge and sea-ice melt/growth anomalies can act either constructively or destructively in the creation of ocean freshwater anomalies.

4 Conclusions

This work represents a comprehensive study of a series of Arctic freshwater anomalies, achieved through a synthesis of observations and models, into the Mackenzie River basin, through the terrestrial system, to their fate in the Beaufort Gyre of the Arctic Ocean. Physical processes such as lake storage and sea-ice melt/growth acted to modify the flows in important ways that might not have had such a strong influence in other regions of the Arctic. To help summarize our findings, Fig. 13 plots monthly anomalies of P–E, river discharge, ocean freshwater content, and the height of Great Slave Lake.

Discharge to the Arctic Ocean from the mouth of the Mackenzie was at a record low for WY 1995. Water-year discharge in 1996, while near average, reflects compensation between small or negative flow anomalies from October to June and above-average flows from July to September. Record high water-year discharge in 1997, with above-average flow each month, was followed by record high flow in May 1998, then a sharp decline. Although the monthly discharge and P–E time series have high intrinsic variability,

annual river discharge anomalies in both WYs 1995 and 1997, in May of 1998, and the P–E anomaly during WY 1996 are statistically significant hydrologic signals.

As Fig. 13 helps to further illustrate, relationships between river discharge anomalies and those of P–E are influenced not only by season-to-season lags, but a suite of within-basin processes. Widely studied as part of MAGS, the low discharge for WY 1995 reflects both negative anomalies in P–E and a pattern in which recycled summer precipitation fell over the southern part of the basin, characterized by low runoff ratios and dry surface conditions immediately prior to the water year. The hydrograph for WY 1996 appears to reflect a combination of spring soil moisture recharge, buffering of spring flows by rising lake levels along the watercourse, positive P–E anomalies in July and August, and a massive summer release of water held in storage by Bennett Dam. The record high water-year discharge of 1997 manifests the dual effects of reduced buffering by lakes and positive P–E anomalies for most of the year. As highlighted in bold in Fig. 13, lake levels remained high, so that even below-average terrestrial P–E the following autumn and winter, associated with a shift in atmospheric circulation linked to the onset of El Niño conditions, resulted in record discharge in May 1998. Lake levels then fell sharply throughout the remainder of the year.

An oceanic freshwater anomaly formed in the Beaufort Sea in autumn 1996, remaining largely stationary through the winter and early autumn of 1997. By autumn 1997, and as indicated in the bottom panel of Fig. 13, the anomaly moved off-shelf and was entrained into the anticyclonic Beaufort Gyre circulation. The anomalously positive river discharge in 1996 contributed approximately 16 to 24% of the oceanic freshwater anomaly, the remainder associated with reduced

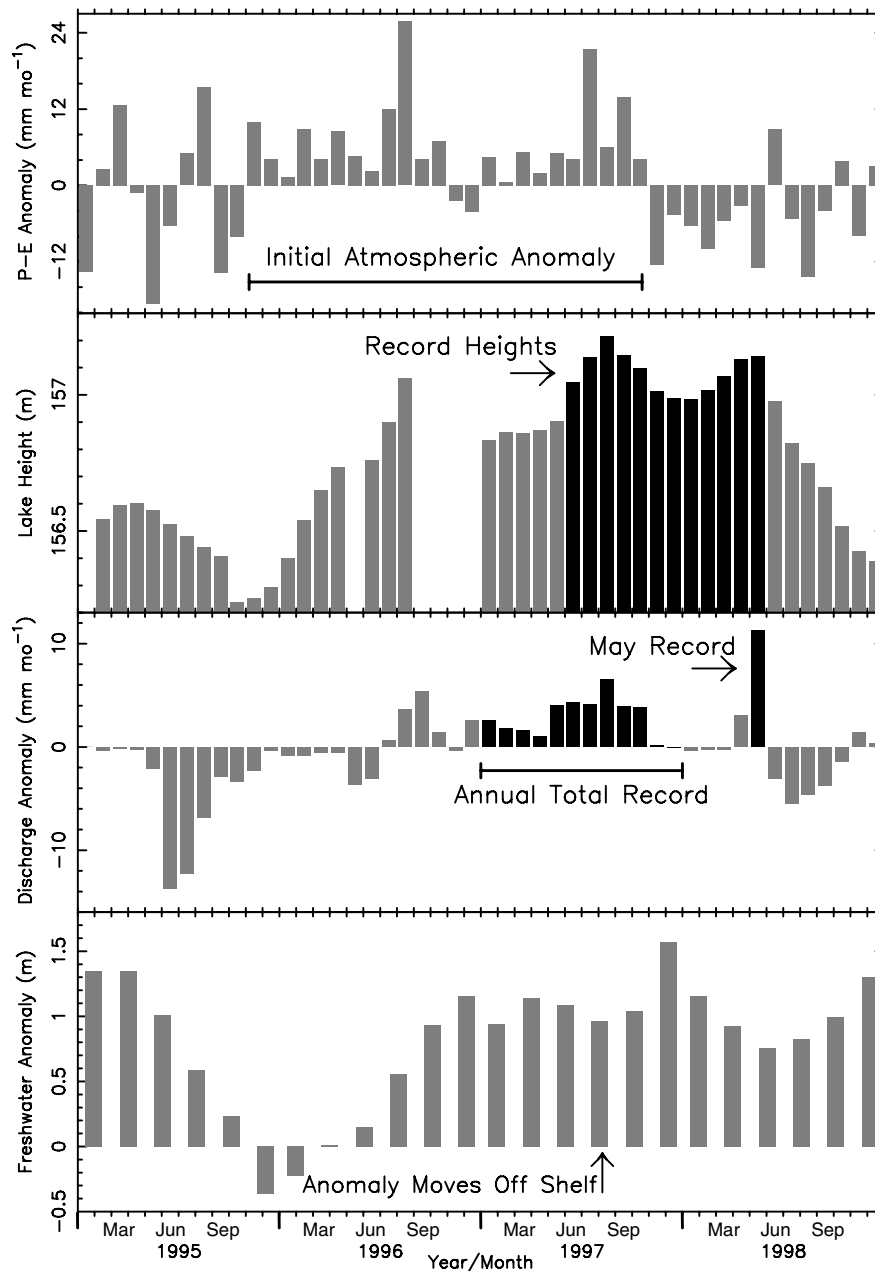


Fig. 13 Anomaly in P-E across the Mackenzie basin (mm mo^{-1}), the height of Great Slave Lake (m), anomaly in river discharge past gauging station at Arctic Red River (mm mo^{-1}), and upper 65 m freshwater anomaly on the Mackenzie shelf and in the Beaufort Gyre during 1995–98 (m). Four months are missing in the lake height record. Freshwater anomalies are computed as an average over two months. From Jan–Feb 1995 to May–Jun 1997 the anomaly magnitude is determined as a spatial average over the Mackenzie shelf (black-outlined area in Fig. 12b). In July 1997 the anomaly moves off-shelf, and the magnitude is computed following the anomaly toward its location in Fig. 10b. It gradually weakens in later years (1999–2002) as shown in Figs 10 and 11.

ice growth due to heavy ice conditions, increased summer ice melt, and to a lesser extent, positive P-E anomalies over the ocean itself. While the excess river discharge contributed to the magnitude of the anomaly, it did not significantly affect its evolution through its effect on ocean circulation. Starting in autumn 1997 and throughout 1998, and despite strong river inputs, the upper ocean actually became more saline. While puzzling at first, it appears that the effects of anomalous river

input were overwhelmed by strong ice growth and associated brine rejection.

It is hoped that the present study, following freshwater anomalies through the atmosphere, the land surface and lakes of the Mackenzie watershed, to their ultimate fate in the Beaufort Sea, serves to highlight both the fascinating complexity of the Arctic freshwater system and the value of synthesizing observations and models.

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