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Abstract

Climate change and population growth is putting increased pressure on water supply. However, detailed water-balance information, which would assist with management is lacking for major reservoirs around the world. This information is particularly critical in mid-latitude northern Mediterranean climates where evaporation is a potentially important water-balance component.

This study examines the seasonality of the water balance for the Sooke Reservoir in western Canada, a major water supply for the City of Victoria, British Columbia. Evaporation is estimated with three evaporation models, Penman, Priestley-Taylor, and Hamon and the results are compared. Inflows are estimated with the contributing-area approach and the HBV-EC, hydrologic model. Finally, a worst-case drought scenario is created.

If conditions of low precipitation and high evaporation like those found in the study period were to persist, water levels would become critically low during the third dry season and by the fifth season if water restrictions were put in place.
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CHAPTER 1: INTRODUCTION

1 BACKGROUND

Renewable sources of fresh water on the earth’s surface are limited and irregularly distributed in space and time. Humans have contained water supply in one location by collecting it creating a more reliable and constant supply despite its natural variation. Reservoirs are replenished by many sources including streamflow, groundwater, snow, and/or rainfall. They are diminished by multiple losses including consumption and evaporation. These inputs and outputs are characteristic of the hydroclimatology of the region (Koeppen and De Long, 1958; Stahl and Hisdal, 2004). Water storage is designed to meet multiple objectives such as hydropower, irrigation, potable supplies, fishing and recreation, and to reduce the risk of floods and droughts (UNESCO, 2006). Although dams and reservoirs have the positive benefit of providing a more reliable supply, they have the negative effect of disrupting natural ecosystem functions (Dynesius and Nilsson, 1994; Prowse et al., 2004; Rosenberg et al., 2000). Construction of new dams is inhibited both by a lack of available space and by strong social pressure against such unnatural interference.

Canada has been ranked second in the world for water availability, yet some communities have been experiencing water supply shortages that were caused by water quantity and/or water quality problems (Sullivan, 2002). From 1994 to 1999, about 26% of the municipalities in Canada with water-supply systems reported water shortages for reasons such as drought, infrastructure problems, and increased consumption (Marsalek et al., 2004). Residential water use in Canada accounts for more than half of all municipal water use and ranges from 240 to 460 litres/capita/day, which are some of the highest rates among the developed nations (Marsalek et al., 2004). As good quality supplies become limited, we are moving to lower quality sources that require more treatment. These may increase the risk of
human health problems by increasing the reliance on technology and safe operation to provide a clean drinking water supply.

Climate change is increasing pressure on existing reservoirs, especially in southern Canada. The effects include reduced flows and levels in rivers and lakes, declining groundwater levels, and higher water temperatures (Marsalek et al., 2004). Lower water quality is expected, with increased suspended solids resulting from more frequent severe storms, increased water use with higher air temperatures, and an effect on water distribution related to increasing bacteria growth (Marsalek et al., 2004). The projected change in air temperatures and precipitation will adjust not only the annual, but also the seasonal hydro-climatic patterns to which management systems have been shaped. Greater changes to air temperatures are expected in winter and spring (Rodenhuis et al., 2007). It is anticipated that winter storms will increase in intensity and that the dry season will become longer in some places (Whitfield and Taylor, 1998). These climatological changes will result in adjustments to the timing of hydrological processes such as streamflow, soil moisture, and evaporation.

Natural modes of climate variability, such as the El Nino Southern Oscillation (ENSO) and the Pacific Decadal Oscillation (PDO), may either amplify or diminish the influence of climate change. These modes or teleconnection patterns may also result from or be intensified by climate change. Many of these teleconnection patterns have been studied so that their influence on different regions is understood and can be used to forecast conditions to assist with management. Applying this approach to a given site often depends on establishing the relationship of these modes to the variable in question over multi-decadal periods.

In spite of the reliance of modern society on these vulnerable systems, limited work has been done to evaluate the water balance on which the storage in a reservoir is dependent. Storage is a result of several inputs and outputs including precipitation, inflows, evaporation, consumption, and spill. Climate change and variability impact these terms individually and
collectively, potentially magnifying effects and challenging those attempting to project future storage volumes. There are few sites in Canada where adequate data have been collected to allow the water balance of the system and its response to climate change and variability to be examined. Given adequate data, it is a high priority that research of this type be focused on the most vulnerable locations. These areas are heavily populated, strongly influenced by climate change and variability, and operate close to their capacity under the current climate.

2 STUDY AREA

The Sooke Reservoir (SR) is ideal to study the effects of climate change and variability on water storage due to its location, data availability, and the ability to build on previous work conducted in this system (Figure 1-1). Located in the most southwestern region of Canada, the reservoir has a northern Mediterranean climate characterized by distinct wet and dry seasons. Close proximity to the Pacific Ocean links its hydroclimatology to teleconnection patterns such as ENSO and PDO (Cayan et al., 2001; Fleming et al., 2007; Kiffney et al., 2002). Analysis of past trend have found increases in air temperature in western Canada that surpass the global average, which are thought to result at least partially from climate change (IPCC, 2007). The current population supplied by the SR is just over 350,000 and by 2026 is expected to reach nearly 450,000 (CRD, 2005). Demand for water from the SR is created not only by domestic, but also by municipal and industrial uses and is exaggerated by rapid population growth.
Figure 1-1 - Sooke Reservoir (SR) and surrounding management boundary enclosing most of the Sooke Catchment. Council Creek Basin from which water can be diverted to the SR is outlined. The adjacent Leech Catchment is shown in the inset, along with the location of SR relative to British Columbia.
3 PREVIOUS STUDIES OF THE SOOKE RESERVOIR

The Sooke Reservoir Catchment (SRC) and the Greater Victoria Water Supply (GVWS) have been investigated by in-house, academic, and private-sector research. An internal Capital Regional District (CRD) assessment of the GVWS system’s vulnerability to climate-induced shortages revealed that the SR is managed at only 96% system reliability (most reservoir-reliant systems are managed at 99%). However, contractually the CRD does not have the right to raise the dam at the SR. The watershed adjoining the SR, the Leech Reservoir, is marked as the future water supply. However, it is a limited solution because it currently has poor water quality due to logging. Also, the Leech watershed is not dammed and low-flows occur at the same time as those in the SR (Kolisnek, 2005). In the future, further pressure will be placed on the system not only from increasing population, but also by greater numbers of property owners wanting to switch from groundwater to a connection with the GVWS (Kolisnek, 2005). This is because the water quantity and quality of groundwater wells in the region are declining and many users want to move to the GVWS because it is considered to be a more reliable source (Kolisnek, 2005).

Miles and Associates Ltd. (1994) estimated total inflows using lake level and daily water withdrawals since 1916. Trends in climate data were also evaluated including a basic Thornthwaite evaluation of the average seasonal variation in the water balance. Results showed that the losses from the SR over the summer months exceeded the volume of water withdrawn for consumption, based on discrepancies in storage volumes. These discrepancies were thought to represent evaporation from the reservoir surface or other unaccounted losses or imprecision, but were not known (Miles and Associates Ltd., 1994). Losses via evaporation were studied by Nowlin et al. (2004) using the Morton estimation method. Total evaporation over the January 2001 to December 2002 period was estimated with this method.
The CRD Water Department runs a Hydrological Simulation Program - FORTRAN (HSPF) model as the operational forecast model for the catchment area (Howard and Associates, 1996). Although this model successfully projects weekly inflow volumes to the entire catchment area, it is run with constant monthly evaporation values based on 30-year climate normals and is lumped by elevation. Testing of the model on sub-catchments, such as Judge Creek, has shown that the model can significantly underestimate outflows. This indicates the model does not represent the individual water-balance components completely, which is necessary to understand the influence of climate change on this supply.

4 RESEARCH PROBLEM

In the past, expansion of the reservoir to increase storage has been the most practical, cost-effective approach to meeting increasing demand resulting from population growth. Recently, there have been occasions where supply was limited and strict demand-side management practices had to be introduced to ensure that supply would last until the start of the next wet season. In response to these events, the dam for the SR was raised to capture more water over the winter months. This upgrade temporarily increased the security of the supply, but it may not be a permanent solution in the light of potential population growth and the effect of climate change. Expansion of the supply to include other water sources, such as the Leech River (Figure 2-1), have been considered to meet projected growth in demand. However, further infrastructure is needed to facilitate the use of water from this source, including an enhanced water treatment facility (CRD, 2006; Kolisnek, 2005). Hence, a better understanding of the SR water balance would be invaluable for managing the current system.

One of the major unknowns in the SR water balance is seasonal evaporation. Given the northern Mediterranean climate, evaporation is suspected to be a large draw from the system during the dry season and could be especially important in drought years. However, little is known about open-water evaporation rates in this region. Drought has occurred here
in the past (Walker, 2002) and will probably occur in the future. Drought conditions in Canada have been shown to be related to warm sea surface temperatures (SSTs) in the Pacific Ocean and increases in SSTs are projected for the future (Shabbar and Skinner, 2004). Furthermore, once droughts begin they are likely to span several years because of the larger-scale forcings that promote them (Shabbar and Skinner, 2004).

Although intensively monitored, compared with other catchments in Canada, a comprehensive accounting of the seasonal water balance of the SR has not been previously undertaken. Specifically, evaporation and total inflows have not been estimated independently from the other terms in the water balance on a monthly or seasonal basis. The water-balance terms responsible for extreme low-water wet and dry seasons have not been identified or used to investigate the cause of drought.

This detailed water-balance analysis has important implications in managing the SR to its full potential and in avoiding critical low-water levels. Furthermore, it will set the foundation for understanding how the water balance of the SR has responded to historical climate shifts and help project how it might respond to changing climate in the future. On a broader scale, results of this research may provide information as to how other reservoirs in similar hydroclimatic regimes in North America might be affected by climate change.

5 RESEARCH OBJECTIVES

This study has two primary research objectives:

1. To estimate evaporation from the SR using a seasonally sensitive model.

2. To close the water balance over a contemporary period with high quality data, i.e. October 1996 to September 2005, at a monthly scale, to investigate its seasonality, and examine a worst-case drought scenario for the system.

Additionally, a prerequisite to this work is to estimate total inflows to the SR. Hence, this will be a sub-objective. A water-balance equation for a reservoir can be written as:
\[ I_S + I_G + P - E - O_S - O_G = \Delta S \pm \varepsilon \]  

where \( I_S \) is surface inflow, \( I_G \) is inflow from groundwater, \( P \) is precipitation onto the reservoir, \( E \) is evaporation from the reservoir, \( O_S \) is surface outflow, \( O_G \) is groundwater outflow, \( \Delta S \) is change in storage, and \( \varepsilon \) is the error accumulated from each term (all in \( m^3 \) month\(^{-1} \)).

Current gauging of \( I_S \) involves the discharge from 42% of the contributing area. The remaining 58% needs to be estimated to close the water balance. A thorough investigation of groundwater is beyond the scope of this study due to lack of information on subsurface conditions. Some treatment of this will be examined in the section on seasonality, but groundwater will not be treated as a major component. Comprehensive records are available for precipitation. Based on these records, it was found that precipitation gauged at the Sooke Dam represents precipitation over the surface of the SR (Fairbairn, 2003). Evaporation has been estimated from the SR over the 2000-2001 period using the empirical Morton equation (Nowlin et al., 2004) and a regional study of precipitation minus evaporation (P-E) has applied the air-temperature-dependent Hamon estimate of evaporation (Nord, 2003). A detailed analysis of evaporation from the SR using a more sophisticated model has not been completed. The measurements of air temperature, water temperature, radiation, humidity, and wind speed are sufficient to apply a complex and likely more seasonally sensitive estimate of evaporation. Surface outflow from the SR is the sum of consumption, spill, and fisheries release. The resulting water balance equation addressed in this study will therefore be:

\[ I_s + I_G + P - E - O_{s1} - O_{s2} - O_{s3} - O_G \pm \Delta \varepsilon = \Delta S \]  

(2)
where $O_{s1}$ is surface outflow via consumption, $O_{s2}$ is surface outflow via spill, and
$O_{s3}$ is surface outflow via fisheries release (all in m$^3$ month$^{-1}$). The accuracy of the water
balance components estimated in this study will be the degree to which $\Delta S$ matches the
observed change in storage $\Delta S_0$, based on the change in volume in the SR. This volume is
derived from a regression between volume and water levels.

The following chapters addressing these objectives are described as follows:

• **Chapter 2** – provides an overview of the study area, its hydroclimatic setting and
  physical characteristics including the climate, hydrologic response, land use, and
  geology. Following this overview, the reservoir history and characteristics are
defined. Detailed background material on the availability of climatic, hydrological,
and limnological data are also provided to justify the selected period of intense
monitoring (PIM). The detailed information on data sources is provided in this
chapter and not repeated in subsequent chapters.

• **Chapter 3** – is a stand-alone manuscript that compares three different techniques for
  estimating evaporation from the SR over a nine-year period (October 1996 to
  September 2005). These include the Penman, the Priestley-Taylor, and Hamon
  methods. A method for converting the temperature-based Hamon estimate to
  replicate estimates from the more complex Penman method is also explored. This is
  undertaken to access whether the simplified Hamon model can be used to evaluate
  conditions outside the detailed study period when climate data are limited to
  temperature and precipitation.

• **Chapter 4** – is a stand-alone manuscript that presents the full water balance of the SR
  for a nine-year study period and evaluates the seasonality of each term. Comparing
two approaches for estimating total inflows into the SR: (a) the contributing area
  approach; and (b) the HBV-EC model approach is a focus of this work. The closure
of the water balance is tested against the measurement of storage change and an investigation of errors on the water-balance terms is provided. The potential effect of increases in extreme conditions (such as low precipitation and high evaporation) is explored in a worst-case scenario of drought that is constructed from the lowest precipitation wet season and highest evaporation dry season that occurred during the study period.

- **Chapter 5** – provides the conclusions and recommendations for further work in lake evaporation and the water balance of the Sooke Reservoir.

As mentioned above, this thesis is written so that *Chapters 3 and 4* are written as stand-alone manuscripts. *Chapter 1* provided the introduction to the overall thesis, *Chapter 2* will be an in-depth description of the study, and *Chapter 5* will serve to summarize all of the work in the thesis. Due to this format, some of the basic information will be repeated, such as the study area, with only slight modifications from chapter to chapter.
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CRD, 2006. CRD Water Services-Sooke Reservoir Project Description, Capital Regional District.


CHAPTER 2: STUDY AREA

1 INTRODUCTION

This chapter defines the regional hydro-climatic setting of the Sooke Reservoir (SR) and provides a detailed description of the inter- and intra-annual variation in air temperature and precipitation over the most recent climate normal period (1971-2000). The physical characteristics, hydrologic response, land use, and geology of the catchment area are also described. Following this, a depiction of the reservoir history and an account of the available climatic, hydrological, and limnological data are presented to justify the selected study period.

2 REGIONAL HYDROCLIMATIC SETTING

The SR has a northern Mediterranean climate, typified by distinct wet and dry seasons, and mild winters, which is classified as a Cs climatic type (Koeppen and De Long, 1958). These conditions arise from alternating patterns of low pressure (cycloonic) and high pressure (anticyclonic) systems over the year (Moore et al., forthcoming 2007). The close proximity to the Pacific Ocean and mid-latitude location makes this region especially susceptible to changes in storm tracks and onshore winds brought on by westerlies that are associated with the low-pressure systems that occur more frequently in winter (Bryson and Hare, 1974). The position of the westerlies is dictated by the jet stream that tends to be situated over the SR region in winter and farther to the north in summer.

Westerlies often carry ample moisture, generating precipitation in this region that is high for this latitude. Due to the northward advancement of pressure systems during summer the Sub-Tropical, North Pacific high pressure shifts from south of the SR northward, encompassing much of the area continuing up from Oregon and Washington (Moore et al.,
forthcoming 2007). High pressure limits precipitation and promotes warmer temperatures (Chilton, 1981). Consequently, most of the precipitation falls between October and April and the remaining months are relatively dry.

The hydroclimatology of the SR is influenced over the inter-annual, annual, decadal and multi-decadal timescales by modes of climate variability. The dominant modes acting on the region include the Pacific North American Pattern (PNA), El Nino-Southern Oscillation (ENSO), and the Pacific Decadal Oscillation (PDO) (Bonsal et al., 2001a; Cayan et al., 1999; Mantua et al., 1997; Moore et al., forthcoming 2007; Shabbar et al., 1997; Stahl et al., 2006). These are large-scale ocean-atmosphere phenomena are sometimes referred to as “teleconnections”.

ENSO events persist for 6 to 18 months and have two modes El Nino (La Nina), which bring warmer (cooler) temperatures and less (more) precipitation to southern British Columbia primarily during the winter and spring (Shabbar et al., 1997). The PDO operates on decadal timescales and has warm and cool phases that produce effects similar to those of El Nino and La Nina, respectively (Mantua et al., 1997). The PNA tends to be in a positive state during the warm PDO and El Nino events, and is characterized by a strong Aleutian Low (Hsieh and Tang, 2001). However, investigating the influence of these “teleconnections” on the hydroclimatology of the SR is not the focus of this work and therefore this discussion rests with this brief description.

Another potentially significant influence on the hydroclimatology of the SR is human-induced climate change. Air temperatures increased more over the 20th century in the Pacific Northwest (0.7 to 0.9°C), which includes BC, Washington, and Oregon than they did on average for the globe (0.5 to 0.6 °C) (Folland et al., 2001; Mote, 2003a). Across the Pacific Northwest, annual precipitation has increased from 13% to 38% during the 20th century, with the interior of Washington and south-eastern BC experiencing the greatest
change (Mote, 2003a). Changes in temperature and precipitation are not occurring uniformly throughout the year and the probability of extremes is increasing (Folland et al., 2001).

Although temperature and precipitation extremes have not been studied exclusively for the SR, they have been investigated in BC or across Canada. In southwestern Canada, the strongest changes occurred in winter and spring, as characterized by fewer days with extreme low temperatures and more days with extreme high temperatures in these seasons over the 1900-1998 period (Bonsal et al., 2001b). In the coastal BC region, comparisons of 1976-1985 versus 1986-1995 (Whitfield, 2001; Whitfield and Cannon, 2000) and 1946-1955 versus 1986-1995 (Whitfield and Taylor, 1998) showed that summers became longer and drier and fall rains arrived later and had greater intensity in the later decade. As a result, the hydrologic summer was noticeably longer in 1986-1995 in comparison to 1946-1955, with the low-flow period starting earlier in May and June and ending later, extending into late September and early October (Whitfield and Taylor, 1998). Furthermore, the work by Whitfield and Taylor (1998) suggested that coastal, rainfall driven watersheds are very sensitive to changes in temperature and precipitation.

3 PHYSICAL CHARACTERISTICS

The SR drainage basin (Figure 2-1) is located at 48° 30' 50" N latitude and 123° 42' 1" W longitude. Covering an area of 70.1 km² (including the 7.1 km² SR), it represents about 91% of the Greater Victoria Water District’s (GVWD) water supply (CRD, 2006). Water is diverted from the Council Creek watershed to supplement the Sooke Reservoir Catchment (SRC) supply, which adds 10 km² of drainage area when in use (Green and Gillie, 1994).
Figure 2-1 - Sooke Reservoir (SR) and surrounding management boundary enclosing most of the Sooke Catchment. Council Creek Basin from which water can be diverted to the SR is outlined. The adjacent Leech Catchment is shown in the inset, along with the location of SR relative to British Columbia.
The climatology of the SR is based on the most recent 30-year climate normal of (1971-2000), developed using the long-term data that has been collected at the SR dam. Annual average air temperatures range from a minimum of 7.9°C in 1985 to a maximum of 10.2 °C in 1998, with an average of 8.8 °C and a standard deviation of 0.6 °C (Figure 2-2 A). Annual total precipitation ranged from 730.6 mm in 1985 to 2317.3 mm in 1990 (Figure 2-2 B). Over this 30-year period, average annual precipitation was 1640.0 with a standard deviation of 371.7 mm.

![Figure 2-2 A) mean annual air temperature anomalies (°C) B) mean annual precipitation anomalies (mm) from 1971-2000 measured at Sooke Dam.](image)

The northern-Mediterranean nature of the SR climate is typified by an October to March wet season, with peak monthly precipitation totals occurring in November (Figure 2-3A). The lowest monthly precipitation occurs in July. Mean monthly temperatures are at their minimum in January with maximum temperatures in July and August (Figure 2-3B). Precipitation is most variable in November and least variable in June. Air temperatures have the largest variability in January and the least variability in June, July, and October.
Figure 2-3 Box and whisker plots of A) air temperature (°C), B) monthly precipitation (mm) at Sooke Dam over the 1971-2000 period. Whisker plots show minimum non-outlier (bottom) and maximum non-outlier (top) values; the dark black line shows the median; the extent of the box depicts the 25th (bottom) and 75th (top) percentiles; circles show outliers.

Both elevation and distance from the ocean influence the amount of precipitation falling on areas within the Sooke catchment (Niemann 1993; Fairbairn 2003). Winds are primarily southerly, produced by the prevailing winter southeasterlies or summer southwesterlies. However, winter outflow winds may cause strong northerlies, which have been linked to increased wave action on the reservoir surface. Additionally, the topography of the catchment tends to funnel wind north south (Green and Gillie 1994).

The hydrologic response of the SR watershed is reflective of the above noted hydro-climatic characteristics. In general, it can be classified as a predominately pluvial regime where early winter (November or December) is typified by a large hydrograph peak generated from winter rains (sometimes augmented by snowmelt at higher elevations) and is followed by consistent rainfall runoff over the remainder of the winter. This wet season often spans October to March or April, and is common to mid-altitude watersheds in the coast mountain region (Fleming et al., 2007; Moore et al., forthcoming 2007). If there is some
winter accumulation of snow, it does not persist long enough to augment flows in the spring or summer. The dry season normally begins in April as rainfall tapers off and air temperature increases. Dry conditions prevail in August and September, characterized by an extended period of minimal rainfall concomitant with higher temperatures.

The management the SR catchment and its vegetation types has an influence on the hydrology of the basin via evapo-transpiration. The catchment area is closed to visitors and maintained with the objective of protecting water quality, preventing forest fires, and conserving wildlife habitat (CRD, 2007). The majority of the land area within the catchment is forested (83%), with the remainder covered by small lakes and wetlands or open spaces (CRD, 2001). A road network allows for maintenance and provides access for logging practices that take place primarily outside the catchment boundary. Although most logging ceased in the early 1990s (CRD, 2006), about 5% of the catchment area is currently being logged (Kolisnek, 2005). The remaining forests range in age from less than 20 years to between 125 and 250 years (CRD, 2001). Douglas fir is the dominant tree species, but Red Cedar, Pine, Arbutus, Alder, and Maple are also present (Green and Gillie, 1994).

The underlying geology of the catchment area is made up of three major bedrock types (Drown, 1991), each having implications for surface water and groundwater transportation. *Wark Gneiss* produces the most rugged terrain of the three; it contains the Rithet Creek valley and the steeply sloped areas on the west side of the SRC. This steep terrain promotes rapid runoff because it reduces infiltration rates and creates an orographic effect that enhances precipitation. *Colquitz Gneiss* forms hilly or hummocky terrain in the Judge Creek watershed and other portions of the eastern watershed where areas are more gently sloping and pocketed, creating wetlands. The *Leech River formation* is made up of predominantly of argillites (Elwell 1983) and occurs on the south end of the SRC, forming a subdued topography characterized by a series of low ridges (Drown 1991).
Surficial materials at higher elevations consist principally of *colluvium veneers* and *morainal till blankets* overlying bedrock. Exposed bedrock exists primarily in the southern portion of the watershed. Fluvial or glacial-fluvial sediments occur in valley bottoms along the Rithet and Judge Creek. A relatively thin soil layer (~1 m thick) covers the majority of the catchment area.

4 RESERVOIR HISTORY AND LIMNOLOGY

The geometric configuration of the SR has changed as a result of an historical series of increases in dam retention heights. The last occurred in 2002 when it was raised 6 m to a height of 186.75 metres above sea level (masl), increasing the holding capacity by 78% (92.7 x 10^6 m^3 from 52.0 x 10^6 m^3). A secondary reservoir, Deception Reservoir, was built at this time, but was separated from the SR by a central core rock-fill dam and is currently not used for water-supply purposes due to its low water quality. Originally, the dam was located roughly 100 m north of the existing dam and had a spillway elevation of 174.20 masl (first built in 1912). It was upgraded in 1970 and raised to 180.75 masl. In 1992, the intake tower was extended to accommodate the anticipated 2002 expansion.

The lake formed by the dam has been classed as oligotrophic and monomictic (Nowlin et al., 2003) with a surface area to overall catchment ratio of 1:10. Minimum water temperatures decline to 4°C for most years, but have been recorded as low as 3°C. Permanent ice cover has not occurred within the short record of water temperature profiles, which have been taken since 1996. In late summer, maximum surface water temperatures reach as high as 28°C. In late spring through early fall, inflows diminish while water consumption increases, causing the SR water level to decline until the rains arrive in late fall (CRD 2005).

The SR has three basins. The largest is farthest north (area 4.34 x 10^6 m^2) and is the deepest (up to 70 m). Second in area (1.26 x 10^6 m^2) and depth (28 m) is the middle basin. The third is the most southern basin with an area of (0.45 x 10^6 m^2) and a maximum depth of
22 m (Spafard, 2002). These values are based on the pre-December 2002 dam height of 180.80 (masl), 6 masl below the current dam. The lake is 6.0 km long and 1.5 km wide at its widest point.

5 DATA HISTORY SUMMARY

Heavily monitored, undeveloped and secured against outside visitors, the Sooke Reservoir Catchment (SRC) has optimum water quality and provides an ideal setting for studying the hydroclimatology and seasonality of the water budget of the SR. Exploring the history of data collected in the SRC will help to demonstrate the availability of data to meet the two primary objectives of this work; (1) to estimate evaporation from the SR using a seasonally sensitive model; and (2) to close the contemporary water balance on a monthly time-scale.

The SRC data set is unique to most reservoirs in Canada, because monitoring started in the early 1900s. Long-term records are available for air temperature, precipitation, spill consumption, and water level (Table 2-1). In the 1990s, multiple meteorological stations were installed increasing the spatial coverage of precipitation and temperature measurements, and adding instruments to monitor wind speed/direction, humidity, and radiation (Table 2-1). Streamflow was observed in the two largest streams (by area and volume) for the first time in the 1990s. Profiling of the lake-water temperature also began at this time. From October 1996 to September 2005, monitoring was at its maximum both spatially and temporally. Hence, this period is referred to as the period of intensive monitoring (PIM) and is used to estimate evaporation and to close the water-balance equation in this study.

Improved monitoring in the 1990s occurred in several areas. In 1992, a diversion was created to draw water from Council Creek into the SR through Trestle Creek (Figure 2-1) at the most south-eastern point of the SR. Outflow measurements became more precise with the installation of a mechanical totalizer in 1992 (Gudavicius, 2006). In 1993, streamflow
measurements began on Judge and Rithet Creek. Hourly measurements of air temperature, precipitation, and humidity started in 1995 below the dam, and hourly wind speed/direction measurements began on top of the intake tower. Lake-water temperatures were measured at four locations in the lake starting in 1996. Measurements of radiation started in 1998 with the installation of a pyranometer (PYR) and a photometric (PAR) on top of the intake tower. After about 2002, outflow was measured by exception roughly every minute (Gudavicius, 2006). Controlled fishery releases started in February 2004 to provide reliable flow downstream of the SR in the Sooke River. They are measured with a magnetic flow meter using the same method described for outflow.

Prior to the 1990s, minimum and maximum air temperatures were measured from 1919 to 1966 near the dam. Precipitation measurements of total rain, snow, and precipitation were made daily using a manual gauge, starting in 1903, near the dam. In 1971, the dam was upgraded and the precipitation gauge was moved to a tower mounted on top of the water intake platform. The manual gauge remained there until May 1998. Prior to about 1970, when the new Sooke Dam was built, outflow was measured using a weir to determine water released at Sooke Dam. Although official records are not available, the water released at Sooke Dam was likely set on a daily basis and did not change throughout the day. Between 1970 and about 1992, outflow was measured throughout the day by a mechanical meter read once per day. Reservoir water-level measurements up to and including 1998 (starting in 1919) were calculated based on one water-level reading taken during the day.

Table 1 provides an overview of how stations were thought to have been moved over the longer period of record (1903-2005), “x” denote the time period when measurements were made with each instrument. Meta-data for measurements made before the 1990s is limited. Therefore, little is known about when gauges were read and where they were located. Note that between 1966 and 1995 no measurements of air temperature were made in the SRC. Another important consideration is that, in the latter part of the record, measurement of spill
took place on a more frequent basis (hourly). Lastly, there are some unexplained gaps in the water-level measurements over the record, some of which were in-filled by the CRD by estimating water levels from consumption and spill amounts. The meta-data regarding these measurements are limited, which inhibits their application. This comparison of the pre- and post-1990s record demonstrates that more measurements, both spatially and temporally, were made in the later period, especially after 1996, and more metadata is available for this later period.
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Table 2-1 Data summary for the Sooke Reservoir Catchment from 1903 to 2005.
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CHAPTER 3: ESTIMATING EVAPORATION FROM THE SOOKÉ RESERVOIR, BC, CANADA

ABSTRACT

Evaporation is a rarely examined yet potentially important component of the water balance in reservoirs. This is especially true for climates that can be water stressed in the dry season, such as northern Mediterranean climates in mid-latitudes. This study attempts to quantify evaporation for the Sooke Reservoir (SR) in western Canada, a major water supply to the City of Victoria, BC based on three models: the detailed Penman and the Priestley-Taylor methods, and the simpler Hamon method over the 1996 to 2005 study period.

This analysis shows that heat storage in the SR has a strong influence on the seasonality of the evaporation. It typically delays peak evaporation by one month from July to August in the summer season, and drives evaporation in the early winter period when net radiation is low. However, heat storage was only accounted for in the Penman and Priestley-Taylor methods.

The Hamon method was also tested for applicability in estimating evaporation for periods when climate data are extremely limited. Values from the Penman method were selected as the standard to which Hamon estimates were compared. The Hamon approach underestimates annual evaporation and does not have the same seasonal timing and magnitudes as the Penman. A technique for converting the Hamon estimates to match those for Penman was explored and provided reasonable estimates for many months in the year. The inter- and intra-annual variation of the Penman was hard to match equally for all months, due to the inter-annual variation in heat storage.
Two of the warmest years in the instrumental record of global surface temperatures since 1850 (1998 and 2005) were part of the study period. The evaporation estimated during these two extremely hot years serves as a surrogate for possible future evaporation in 2050s as they are similar to the projected temperatures for that period.

1 INTRODUCTION

The Sooke Reservoir (SR) is operated by the Capital Regional District (CRD) to supply water to the Greater Victoria Area (GVA) by capturing and storing precipitation and runoff to meet the demand of the steadily growing population of this urban centre. On several recent occasions, the SR supply has been insufficient to meet demand, suggesting that the long-term net surplus may be diminishing. Over the past century, in the southwest portion of BC (Whitfield and Taylor, 1998; Whitfield, 2001) and over the Pacific Northwest (Regonda et al., 2005), the wet season has become warmer and wetter with larger precipitation amounts falling over a shorter period. The dry season has, in turn, become longer. These changes have been more substantial in the last ten years of the 20th century and have continued into the beginning of the 21st century.

Changes in temperature and precipitation resulting from climate change and variability alter hydrology, which impacts water supplies in BC. Further increases in temperature and changes to precipitation are projected to occur in the future because of climate change (Whitfield et al., 2002). Climate variability can amplify or diminish the effects of climate change, but is also thought to be modified by climate change. This region is known to be strongly affected by two climate variability patterns, the El Nino-Southern Oscillation (ENSO) (Shabbar et al., 1997; Cayan et al., 1999) and the Pacific Decadal Oscillation (PDO) (Mantua et al., 1997). Historically, these patterns have contributed to extreme conditions in this region, such as drought and floods. Understanding the relationship
of evaporation to these factors is becoming increasingly important to allow better management of water supply.

Open-water evaporation is thought to be a key water-balance component for the SR due to limited precipitation amounts during the dry season, warm summer temperatures, year-round ice-free conditions, and significant influxes of net radiation from May through to October. To manage this supply, an estimate of evaporation is needed, particularly to improve the approximation of available supplies over the dry season. Regional estimates of lake evaporation are outdated, often based on pan evaporation approaches and presented as per annum values that prevent understanding their seasonality. Reference is most often made to Ferguson et al. (1970), but this study provides only annual climatological averages of lake evaporation and does not deal with potential changes related to post-1970s climate trends. Therefore, such data sources require updating. For example, Phillips (1990) reported evaporation amounts for pre-1990, but the 12 hottest years on record are known to have occurred post-1994 globally (IPCC, 2007). Furthermore, warming in western Canada in the past century has exceeded that for the globe (IPCC, 2007).

Although several recent evaporation studies from individual lakes have been carried out in North America, estimates in the northern Mediterranean hydroclimatic regime, such as the SR, are largely unavailable. An important difference with many of the studies is that they are at sites where the water body freezes over during the winter, unlike the SR, which has never been permanently ice-covered over the winter season. Evaporation from Okanagan Lake, in BC’s southern interior, was studied in 1980 and 1981. Most of the work in this study was on testing techniques rather than providing estimates to which results from other studies could be compared (Trivett, 1983). Many of the latest Canadian studies were conducted in the high Arctic by Gibson et al., (1996), Rouse et al., (2003) and Blanken et al., (2000a,b) in areas that have an even shorter ice-free season. Other studies, carried out in the USA by Winter et al., (1995), Winter et al., (2003), and Lenters et al., (2005), are on central
continental, mid-latitudes lakes, which have a limited ice free period, commonly spanning only from March to November. Two notable studies conducted in arid climates with mild winters include those of Lake Mead in Arizona and Nevada (Westenburg et al., 2006) and of Klamath and Tule Lakes National Wildlife Refuges in Oregon and California (Risley et al., 2006). In these locations, winters were ice free, but because of the more southern latitudes more net radiation was available during the shoulder seasons than in the SR. Hence, an obvious spatial analogue of evaporation in the SR does not exist.

In the summer and early fall, when water levels are characteristically low, the SR is potentially vulnerable to evaporation. However, the degree to which evaporation has contributed to low-water levels in the past is not understood. Previous water balance work conducted for the SR has suggested that losses of water though evaporation could exceed amounts lost to consumption over the summer period (Miles et al., 1991). Evaporation from the SR has been estimated for only two consecutive dry seasons (Nowlin et al., 2003) with the empirical Morton equation. Estimating evaporation amounts under a variety of conditions and exploring the relationship of evaporation to changes in water level in the SR are important steps towards projecting future water volumes. Robust estimates should be based on the highest-quality data and most detailed metadata.

The purpose of this study is to compare two estimates of evaporation that use reliable, data intensive techniques, the Penman and Priestley-Taylor methods, to each other. And to apply the better of these two techniques to evaluate a method developed originally by Hamon, which is less data intensive. Data have the highest spatial and temporal resolutions over a nine-year period starting October 1996 and ending September 2005. This is referred to as the period of intense monitoring or PIM and has measurements of air temperature, water temperature, relative humidity, net radiation, and wind speed. For 1916 to 1966, only air temperature records are available. The Hamon technique could be used during this data
sparse period and could also assist in estimating future evaporation based on projected air
temperatures from Regional (RCMs) or Global Climate Models (GCMs).

Evaporation is difficult to estimate precisely because it is driven by a multitude of
factors such as air temperature, wind speed, humidity, water availability, water temperature,
and net radiation. It is also challenging to measure directly. The energy budget and eddy
covariance methods of estimating open-water evaporation are considered to be most
representative of true evaporation rates, but they require costly instrumentation and a large
time commitment for data collection and processing, which makes them impractical for some
studies (Winter et al., 1995; Winter et al., 2003; Rosenberry et al., 2007). More than 30
equations are available to estimate evaporation (Winter et al., 1995), which can be
categorized into four main types. These include temperature-based, humidity-based, mass-
transfer and radiation-based (Xu et al., 1998). The accuracy of a given method is dependent
on timescale (Xu et al., 1998, Lenters et al., 2005) with comparable values often found on an
annual basis, but greater differences on seasonal, monthly or weekly timescales.

In a wide range of studies in a diverse array of hydroclimatic regimes, the Penman
approach (Penman, 1948) provides monthly estimates of evaporation that most closely match
those measured in full energy budget approaches (Winter et al., 1995; Valiantzas, 2006;
Linacre, 1993; Shuttleworth, 1993). This technique is sometimes referred to as the Penman
Combination approach because it combines the energy required to sustain evaporation (the
energy budget component) and an empirical description of the diffusion mechanism by which
energy is removed from the surface as water vapour (the mass-transfer or aerodynamic
component) (Shuttleworth, 1993). Modified forms of the Penman that have been developed
for application where only standard weather station data is available give reasonable results
(Linacre, 1993; Price, 1994; Winter et al., 1995; Valiantzas, 2006). Open-water evaporation
estimates may include a heat-storage component that is best estimated by measurements of
the water body itself (Winter et al., 1995; Gibson et al., 1996; Marsh and Bigras, 1988). This
component is especially important for medium-sized lakes at mid latitudes, which can store significant amounts of heat (Derecki, 1975; USGS, 1954). Adequate data have been collected at the SR to facilitate the application of a simplified version of the Penman equation to estimate evaporation over the nine-year PIM that includes heat storage (Werner, 2007a). It is recommended that when applying the Penman equation, wind speeds are measured at 2 m above the water surface. However, the wind speed measurements that are available from the CRD record were measured at varying heights, which were greater than 2 m.

The Priestley-Taylor method is also a highly regarded, but it does not require wind speed. Hence, estimates from Priestley-Taylor will serve to check that wind-speed measurements are not biasing Penman results.

The objectives of this study are:

1. To estimate monthly evaporation using the Penman combination equation over October 1996 to September 2005, the period of intense monitoring (the PIM).
2. To compare the Penman estimates against those obtained with the Priestley-Taylor method.

Both methods require values of net radiation and heat storage. Although shortwave radiation was available for the majority of the record, there were extended periods where it was not known. The CRD archive includes water temperature measurements for the SR that were made on a routine basis, which provided ample data for estimating heat storage. Therefore, estimating shortwave radiation for periods where data are missing and calculating heat storage in the SR will be sub-objectives of this work. The final objective of this work is:

3. To compare a commonly applied air-temperature-based estimate, the Hamon method, to the Penman method and develop an adjustment factor to convert Hamon estimates into more representative values for the Sooke Reservoir.
2 STUDY AREA

The SR drainage basin (Figure 3-1) is located at 48° 30' 50" N latitude and 123° 42' 1" W longitude. Covering an area of 70.1 km² (including the 7.1 km² SR) it represents about 91% of the Greater Victoria Water District’s (GVWD) water supply (CRD, 2006). Water is diverted from the Council Creek watershed to supplement the Sooke Reservoir Catchment (SRC) supply, which adds 10 km² of drainage area when in use (Green and Gillie, 1994). The SR has a northern Mediterranean climate, typified by distinct wet and dry seasons, and mild winters, which is classified as a Cs climatic type (Koeppen and De Long, 1958). These conditions arise from alternating patterns of low pressure (cyclonic) and high-pressure (anticyclonic) systems during the year (Moore et al., forthcoming 2007). The close proximity to the Pacific Ocean and mid-latitude location makes this region susceptible to changes in storm tracks and onshore winds brought on by westerlies (low-pressure systems) (Bryson and Hare, 1974). High pressure limits precipitation and promotes warmer temperatures in the summer (Chilton, 2000). Consequently, most of the precipitation falls between October and April in this region and the remaining months are relatively dry.

The hydrologic response of the SR watershed reflects the above-noted hydro-climatic characteristics. In general, it can be classified as a predominately pluvial regime where early winter (November or December) is typified by a large hydrograph peak generated from winter rains (sometimes augmented by snowmelt at higher elevations) and is followed by consistent rainfall runoff over the remainder of the winter. This wet season often spans November through to March or April, and is common to mid-altitude watersheds in the coast mountain region (Moore et al., forthcoming 2007; Fleming et al., 2007). If there is some winter accumulation of snow, it does not persist long enough to augment flows in the spring or summer. The dry season begins sometime in April as rainfall tapers and air temperature
begins to increase. Dry conditions prevail in August and September characterized by an extended period of minimal rainfall concomitant with higher temperatures.

Figure 3-1 - Sooke Reservoir (SR) and surrounding management boundary enclosing most of the Sooke Catchment. Council Creek Basin from which water can be diverted to the SR is outlined. The adjacent Leech Catchment is shown in the inset, along with the location of SR relative to British Columbia.
The geometric configuration of the SR has changed from a historical series of increases in dam retention heights, the second and last occurring in 2002 when it was raised 6 m to a height of 186.75 masl increasing the holding capacity by 78% (92.7 x 10^6 m^3 from 52.0 x 10^6 m^3). Originally, the dam was located roughly 100 m north of the existing dam and had a spillway elevation of 174.20 masl (first built in 1912). It was upgraded in 1970 and raised to 180.75 masl.

The lake formed by the dam is oligotrophic and monomictic (Nowlin et al., 2003), with a surface area to overall catchment ratio of 1:10. Minimum water temperatures decline to 4°C for most years, but have been recorded as low as 3°C. Permanent ice cover has never occurred, based on water temperature profiles taken since 1996. In late summer, maximum surface water temperatures reach as high as 28°C. In late spring through early fall, inflows diminish while water consumption increases causing the SR water level to decline until the rains arrive in late fall (CRD, 2004).

The SR has three basins. The largest basin is farthest north (area 4.34 x 10^6 m^2) and is the deepest (up to 70 m). Second in area (1.26 x 10^6 m^2) and depth (28 m) is the middle basin and third is the most southern basin with an area of (0.45 x 10^6 m^2) and a maximum depth of 22 m (Spafard, 2002). These values are based on the pre-December 2002 dam height of 180.80 masl, 6 masl below the current dam. The lake is 6.0 km long and 1.5 km wide at its widest point.

3 METHODOLOGY AND DATA

Estimating evaporation by the Penman and Priestley-Taylor methods requires air temperature, humidity, wind speed, radiation, and water temperature. The availability of these data will be outlined in the following section along with the relevant equations. Following this, there will be a description of those components that will be derived from the observed
data via computations to meet the data requirements of the two detailed estimates. Namely, the methodology for deriving heat storage and shortwave radiation will be described. This section will conclude by describing the Hamon method and the methodology used to adjust the first estimates from the Hamon method into values that more closely match those estimated via the Penman method.

### 3.1 Observational Data

Air temperature is measured via a thermistor (accuracy $\pm 0.2^\circ C$) that was installed below the dam at approximately 172.0 masl on November 14, 1995. Measurements are logged every hour on the hour. Humidity measurements started on November 14, 1995 at the toe of the dam (172.0 masl) with an FTS, THS-1 humidity sensor. This instrument is accurate to $\pm 2\%$ between zero and 80% humidity, and to $\pm 5\%$ between 80 and 100% and takes measurements and logs them once per hour. Incoming shortwave radiation is measured to $\pm 5\%$ using a pyranometer (PYR, Licor, model LI-200SZ) that was installed on June 30, 1998 on top of the intake tower at 195.5 masl. The recorded hourly values are the average of measurements taken every 10 minutes during the previous hour.

Wind speed and direction are measured at 197.5 masl on the intake tower. Wind speed is evaluated to within $\pm 0.4 \text{ m s}^{-1}$ using a MetOne WS-013-30. Wind direction is gauged to the nearest degree using a MetOne WD-023-30. From November 13, 1995 to June 30, 1998, measurements were taken only in the last 10 minutes of every hour. Following this, they have been taken every minute and averaged to hourly values. The tower is located approximately 10 m from the southern end of the lake and connected to the edge by a concrete platform. This instrument is positioned at a fixed elevation such that the distance to the water surface has varied from 4 to 16 m in the past depending on the reservoir water level.

Lake-water temperature was measured with a Sea-Bird SBE 19, SEACAT Profiler v3.1 SN 2159 at four locations starting on September 25, 1996 and ending on June 13, 2005.
Water temperature measurement sites include SOL_00, SOL_01, SOL_03, and SOL_04, which are beside the outtake tower, and in the centre of the south basin, the middle basin, and the north basin, respectively (Figure 3-1). Profiles were taken every 3 to 56 days. In total, 278 thermal surveys were undertaken between 1996 and 2005. Due to the importance of heat-storage information in the detailed evaporation estimates, and the irregular nature of measurements estimates were made for the periods that started and stopped on the dates when water temperature profiles were made. These periods are referred to as the Thermal Budget Periods (TBPs) and are discussed later in this section.

Reservoir water-level measurements up to and including 1998 were calculated on a daily water-level reading whose timing varied over the record. Starting on December 6, 1998, hourly readings were taken with an automated water-level recording device (Pers. Comm. Gudavicius 2006). Reservoir surface area and volume were based on relationships between water-level and surface area, and water-level and volume, respectively, which span the range from a water-level of 113.6 m to 190.0 m, as provided by the CRD.

### 3.2 The Penman Estimate

Penman (1948; 1963) described evaporation from open water using (Shuttleworth, 1993):

\[
E_p = \frac{s}{s + \gamma} \cdot \frac{(R_n - Q_{HS})}{\lambda} + \frac{\gamma}{s + \gamma} \cdot \frac{6.43(f_u)D}{\lambda} \quad (\text{mm d}^{-1})
\]  

where \(E_p\) is potential open water evaporation or evapo-transpiration; \(R_n\) is net radiation at the surface (MJ m\(^{-2}\)d\(^{-1}\)); \(Q_{HS}\) is the heat flux into the lake (MJ m\(^{-2}\)d\(^{-1}\)); \(s\) is the slope of the saturation vapor pressure curve (kPa °C\(^{-1}\)); \(\gamma\) is the psychrometric coefficient (kPa °C\(^{-1}\)); \(\lambda\) is the latent heat of vaporization (MJ kg\(^{-1}\)); \(f_u\) is the wind function (m s\(^{-1}\)); and \(D = (e_s - e_a)\).
– $e_a$ is the vapour pressure deficit (kPa) where $e_s$ is saturation vapour pressure (kPa) and $e_a$ is actual vapour pressure (kPa).

All variables in equation (1) can be derived from the climate records of the SR in the PIM, which include hourly values of air temperature (°C), shortwave radiation (MJ m$^{-2}$d$^{-1}$), relative humidity (%), and wind speed (m s$^{-1}$). From these data the Penman equation can be formulated for routine hydrologic engineering applications, which are based on those outlined by Shuttleworth (1993). Vapour-pressure deficit and wind speed are necessary components of the Penman method.

Because the vapour-pressure deficit is not a linear function of air temperature, the following procedure (Shuttleworth, 1993), is used to compute its mean value:

$$D = e_{S(\text{av})} \left(1 - \frac{RH}{100}\right) \text{ (kPa)} \quad (2)$$

where $RH$ is the relative humidity (%) and $e_{S(\text{av})}$ (kPa) is mean saturation vapour pressure for the examined time interval estimated as:

$$e_{S(\text{av})} = 0.5\left[e_s(T_{\text{max}}) + e_s(T_{\text{min}})\right] \text{ (kPa)} \quad (3)$$

$e_s(T_{\text{max}})$ (kPa) and $e_s(T_{\text{min}})$ (kPa) are saturated vapour pressures corresponding to daily $T_{\text{max}}$ (°C) and $T_{\text{min}}$ (°C), respectively. The saturated vapour pressure $e_s(T)$ (kPa) at a given air temperature $T$ (°C) is computed using:

$$e_s(T) = 0.611 \cdot \exp\left\{\frac{17.27 \cdot T}{T + 237.2}\right\} \text{ (kPa)} \quad (4)$$
The wind function \( f_u \) was originally proposed by Penman (1948, 1963) as:

\[
f_u = a_u + b_u U_2 \quad \text{(m s}^{-1} \text{)} \quad (5)
\]

where \( a_u \) and \( b_u \) are wind function coefficients and \( U_2 \) is wind speed at 2 m height (m s\(^{-1}\)). In the original Penman (1948, 1963) equation \( a_u = 1 \) and \( b_u = 0.536 \). This version of the equation will be referred to as \( f_u^{(1)} \). Penman suggested the \( a_u \) be revised to 0.5 in 1956, which resulted in \( f_u^{(2)} = 0.5 + 0.536 U_2 \). \( f_u^{(1)} \) overestimates the evaporation from large open-water surfaces because the aerodynamic resistance \( (r_a) \), which is inversely proportional to \( f_u \), increases for much larger expanses of water thereby reducing the evaporation rate (Brutsaert, 1982; Cohen et al. 2002; Valiantzas, 2006).

Wind speeds were measured at variable heights by the CRD and were adjusted to 2 m using the equations suggested by (UNESCO-IHE, 2007):

\[
U_a = \frac{2.11 U_h}{\log(66.7 \cdot h - 5.3)} \quad \text{(m s}^{-1}\text{)} \quad (6)
\]

where \( U_a \) (m s\(^{-1}\)) is the adjusted wind speed and \( U_h \) (m s\(^{-1}\)) is the wind speed at height \( h \) (m).

### 3.3 Priestley-Taylor Estimate

The Priestley-Taylor estimate for evaporation is given by:

\[
E_{PT} = u_0 \left( \frac{s}{s + \gamma} \right) \cdot \frac{R_s - Q_{hs}}{\lambda} \quad \text{(mm d}^{-1}\text{)} \quad (7)
\]
where $\nu = 1.26$ (dimensionless) (Stewart and Rouse, 1977). Testing on various surfaces, with unlimited water supplies, the overall mean of 1.26 for $\nu$ was found and considered to represent potential evaporation for humid locations (Priestley and Taylor, 1972; Stewart and Rouse, 1976). Values were converted into (mm month$^{-1}$) as described for the Penman equation.

The following approximations are also used in both the Penman Eq. (1) and Priestley-Taylor computations:

$$\lambda = 2.501 - (2.361 \times 10^{-3}) \cdot T$$ (MJ kg$^{-1}$) \hspace{1cm} (8)

and

$$s = \frac{4098 \cdot e_s}{(T + 237.3)^2}$$ (kPa °C$^{-1}$) \hspace{1cm} (9)

where $s$ is the slope of the saturation vapour pressure curve and

$$\gamma = 0.0016286 \cdot \frac{\Pi}{\lambda}$$ (kPa °C$^{-1}$) \hspace{1cm} (10)

is the psychometric constant where $\Pi$ is the atmospheric pressure estimated by the following equation:

$$\Pi = 101.3 \cdot \left( \frac{293 - 0.0065 \cdot Z}{293} \right)^{5.26}$$ (kPa) \hspace{1cm} (11)
where $Z$ is masl of the location (186 masl).

$R_n$ is computed as the difference between the incoming net short wave radiation, $R_{nS}$, and the outgoing net long wave radiation, $R_{nL}$:

$$R_n = R_{nS} - R_{nL} \quad (MJ \ m^{-2} \ d^{-1}) \quad (12)$$

The $R_{nS}$ is calculated as:

$$R_{nS} = (1 - \alpha) \cdot R_S \quad (MJ \ m^{-2} \ d^{-1}) \quad (13)$$

where $R_S$ is the measured or estimated incoming solar radiation; $\alpha$ is the reflection coefficient or albedo. A typical value of the albedo for open water surfaces $\alpha = 0.08$ (Shuttleworth, 1993; Allen et al., 1998).

In cases when shortwave radiation $R_S$ was not measured or was unreliable, it was calculated with the Bristow and Campbell (1984) method using minimum and maximum daily air temperatures measured at the SR. Calculations were required for October 1, 1996 to July 1, 1998 because the PYR sensor, which measures incoming solar radiation, was not installed until July 1, 1998, well after the other meteorological instruments had been installed on November 14, 1995. Additionally, calculations were made for the June 13, 2002 to March 8, 2003 because the readings from the PYR sensor for this period were judged to be unreliable by the CRD staff.

Using the Bristow and Campbell (1984) technique, shortwave radiation $R_S$ was calculated in the following way:

$$R_S = T_i \cdot Q_o \quad (MJ \ m^{-2} \ d^{-1}) \quad (14)$$
where $T_t$ is the daily total atmospheric transmittance (dimensionless) and $Q_o$ is the daily extraterrestrial insolation incident on a horizontal surface ($J \, m^{-2}$). Daily total atmospheric transmittance is calculated as:

$$T_t = A \left[1 - \exp(-B \Delta T^C)\right] \quad \text{(dimensionless)} \quad (15)$$

where $\Delta T$ is the daily range of air temperature ($^\circ C$), and $A$, $B$, and $C$ are empirical coefficients (dimensionless) determined for a particular location from measured solar radiation data. Daily extraterrestrial insolation ($Q_o$) incident on a horizontal surface ($J \, m^{-2}$) was computed according to Gates (1980) (see Bristow et al., 1984) as:

$$Q_o = 86400 S_o \left(\frac{d}{\bar{d}}\right)^2 \left(h_s \sin \phi \sin \delta + \cos \phi \cos \delta \sin h_s\right) / \pi \quad (J \, m^{-2}) \quad (16)$$

where $S_o$ is the solar constant (1360 W m$^{-2}$), $\bar{d}$ is the mean value of the distance from sun to earth (m), $d$ is the estimated distance from sun to earth (m), $h_s$ is the half day length ($\cos h_s = -\tan \phi \tan \delta$; radians), $\phi$ is the latitude of the location (radians), and $\delta$ is the solar declination (radians). $\left(\frac{d}{\bar{d}}\right)^2$ was taken as unity because it does not differ from unity by more than 3.5% (Gates, 1982 in Bristow et al., 1984).

This method accounted for 70-90% of the variation in daily solar radiation when tested on three data sets in the Pacific Northwest (Bristow and Campbell, 1984). This relationship was investigated on a monthly basis by plotting daily $\Delta T$ ($^\circ C$) versus the daily transmission coefficient. SOLVER (Excel) was used to estimate coefficients $A$, $B$, and $C$ in Eq. (15) to provide daily total atmospheric transmittance values that matched measured
shortwave radiation data. The regression between daily total atmospheric transmittance estimated with Eq. (15) and measured shortwave radiation had a Pearson’s r of 0.94 suggesting a robust relationship. This technique was therefore used to estimate missing values on monthly scales.

\[ R_{nl} \text{ is computed as:} \]

\[ R_{nl} = f\varepsilon' \sigma \cdot (T + 273.2)^4 \quad (\text{MJ m}^{-2} \text{d}^{-1}) \quad (17) \]

where \( R_{nl} \) is outgoing net long wave radiation; \( f \) is adjustment for cloud cover; \( \varepsilon' \) is net emissivity between the atmosphere and the ground; \( \sigma \) is Stephan-Boltzmann constant = 4.903 x 10^{-9} (MJ m^{-2} oK^{-4} d^{-1}); \( T \) is mean air temperature for the examined time interval (°C).

The parameter \( f \) is computed as:

\[ f = \left( 1.35 \frac{R_{ns}}{R_{SO}} - 0.35 \right) \quad (\text{dimensionless}) \quad (18) \]

where \( R_{SO} \) is clear sky radiation MJ m^{-2} d^{-1} (Shuttleworth, 1993) given as:

\[ R_{SO} = (0.25 + 0.5 \cdot 1) \cdot R_A \quad (\text{MJ m}^{-2} \text{d}^{-1}) \quad (19) \]

\( R_A \) is calculated as:

\[ R_A = 15.392 \cdot d_r \left[ \omega_S \sin(\phi) \sin(\delta) + \sin(\omega_S) \cos(\phi) \cos(\delta) \right] \quad (\text{MJ m}^{-2} \text{d}^{-1}) \quad (20) \]
where $\phi$ is the latitude of the location of interest (radians) and $d_r$ is the relative distance between the earth and the sun given by:

$$d_r = 1 + 0.033 \cos \left( \frac{2\pi J}{365} \right) \quad \text{(m)} \quad (21)$$

where $J$ is the Julian date. The solar declination, $\delta$ is given by:

$$\delta = 0.409 \sin \left( \frac{2\pi J - 1.405}{365} \right) \quad \text{(radians)} \quad (22)$$

and $\omega_S$ is the sunset hour angle given by:

$$\omega_S = \arccos(-\tan \phi \tan \delta) \quad \text{(radians)} \quad (23)$$

The $\varepsilon'$ term is calculated as:

$$\varepsilon' = (0.34 - 0.14 \sqrt{e_a}) \quad \text{(dimensionless)} \quad (24)$$

where $e_a$ is vapour pressure is estimated from:

$$e_a = \frac{RH}{100} e_s(T) \quad \text{(kPa)} \quad (25)$$
And the saturated vapour pressure $e_s(T)$ at air temperature $T$ is computed with equation (3).

Other more exact formulations of long-wave energy exchange have been suggested by Brusaert (1982), Croley (1989) and Dingman (1994), but they are not incorporated due to the requirement of temperature at the soil and water surface that are not available for the SR catchment.

The mean change of heat storage $Q_{HS}$ is calculated from:

$$Q_{HS} = 0.0864 \cdot \frac{\rho_w c_w}{a_s} \sum_z \left( \frac{\Delta T_w(z)}{\Delta t} a(z) \Delta z \right) \text{ (MJ m}^{-2} \text{ d}^{-1}) $$(26)

where $\rho_w$ is the density of water 1000 kg m$^{-3}$, $c_w$ is the heat capacity of water $4.19 \times 10^3 \text{ (J kg}^{-1} \text{ °C}^{-1})$, $a_s$ is the mean lake surface area (m$^2$), $\Delta T_w(z) = T_{w2}(z) - T_{w1}(z)$ is the change in water temperature between the second day (°C) and the first day at depth $z$ (m), $\Delta t$ is the number of days between measuring intervals (converted to seconds), $a(z)$ is the lake area at depth $z$ (m), and $\Delta z$ is the layer thickness (m; typically 1 m). Lake area is calculated from a regression of surface area (generated from a bathymetric map) to water level, and water-level measurements are taken daily.

On a given day, the available temperature profiles in the northern, intermediate and southern basins were weighted by the percentage area they represented (Spafard et al., 2002). On days when SOL_04, SOL_03, and SOL_01 were all available, they were weighted by 0.5741, 0.3512, and 0.0747 respectively (four decimal places were maintained to avoid the effect of rounding errors on the heat storage computations). Often only SOL_04 and SOL_01 were available. In these cases, they were weighted by 0.9253 and 0.0747, respectively. On some days when the other profiles were not available, SOL_04 was used to represent the
temperature profile for the entire lake. This profile is the only one of the four profiles that represents the deepest depths at between 22.4 and 70.5 m below the surface. It is also representative of the largest basin in the lake by volume.

Due to the importance of heat-storage information in the detailed evaporation estimates, and the irregular times at which measurements were made, estimates of heat-storage flux were made for periods that started and ended on the dates when water temperature profiles were made. These periods are referred to as the Thermal Budget Periods (TBPs). After estimates were made for the TBPs they were converted into monthly values (mm month\(^{-1}\)). This was done by multiplying the evaporation rate over a given TBP by the number of days in the period. Resulting values were pro-rated by the number of days each period was within a given month and summed to get an estimate for each month.

### 3.4 The Hamon Estimate and an Adjustment Factor

The equation proposed by Hamon for estimating evaporation (\(E_{H}\)) is:

\[
E_H = 2.98 \cdot L \cdot \frac{e_s(T)}{T + 273.2} \quad (\text{mm d}^{-1}) \tag{27}
\]

where \(L\) is the day length (hours; Hamon, 1961; Hamon, 1963; Dingman, 1994). Values were converted into (mm month\(^{-1}\)) by summing the depth (mm) of evaporation for each day in the month.

The Hamon method can be applied when air temperature is the only meteorological record available. Heat storage is thought to have a large influence on the timing and magnitude of evaporation from medium-sized lakes in mid-latitudes, however, this method does not accounted for it. Hence, estimates of evaporation employing the simple Hamon
method are not expected to have the seasonal sensitivity of the more detailed Penman method, which does incorporate heat storage.

A method was devised to adjust Hamon estimates to better match those from the Penman. This technique could be used to estimate evaporation outside the detailed study period when climate data are limited. For this method, Penman evaporation values were divided by those for Hamon to get the percentage ratio (dimensionless) between the two estimates for each month in the PIM. These values were averaged by month to create a monthly conversion factor, $\psi_m$ (dimensionless), where $m$ is the month. Monthly Hamon values were then multiplied by $\psi_m$ to obtain an adjusted Hamon estimate for the Sooke Reservoir that better accounted for heat storage, $E_{HSC}$.

4 RESULTS AND DISCUSSION

Results of the detailed Penman, Priestley-Taylor, and simplified Hamon estimates of evaporation are presented in Tables 3-1 to 3-6 and Figures 3-1 to 3-10.

4.1 The Penman Estimate

Evaporation was calculated initially based on TBPs, an approach restricted by the availability of thermal surveys (see Section 3.3). The Penman method predicts a wide range of variability in evaporation (Figure 3-2 - E). The highest rate (7.7 mm d$^{-1}$) occurred over the August 5$^{th}$ to 11$^{th}$, 1998 or TBP 685. The lowest (-0.6 mm d$^{-1}$) was observed from February 12$^{th}$ to 20$^{th}$, 1997 or TBP 16. The average for the study period was 2.3 mm d$^{-1}$ and the standard deviation was 1.9 mm d$^{-1}$.

A clear seasonal cycle of evaporation is evident. Low rates occur in November to March and tend to increase in May or June and peak in July or August. In late August or September, they start to decrease reaching their minimum around November. In Figure 3-2, evaporation rates are shown to change quickly, likely in response to short-term weather
events. For example, rates dropped from 7.7 mm d\textsuperscript{-1} (5\textsuperscript{th}-11\textsuperscript{th} August 1998, TBP 66) to 5.7 mm d\textsuperscript{-1} (20\textsuperscript{th}-25\textsuperscript{th} August 1998, TBP 699), which is a 34% decrease in less than one month.

Considering this high inter- and intra-monthly variance, long-term averages of monthly evaporation (i.e., 10-30 years) are not likely to be representative of evaporation for a given month, or for periods within a month. This was also observed by Sturrock et al. (1992), Winter et al. (2003), and Lenters et al. (2005) from energy-budget studies of lakes in the mid-continental United States.

The evaporation rates over the TBPs were weighted and adjusted to depict representative monthly values (Section 3.3; Figure 3-3 – A). The average monthly evaporation was 60.8 mm with a standard deviation of 51.1 mm and coefficient of variation of 0.84 over the study period (Table 3-1).

<table>
<thead>
<tr>
<th>Month</th>
<th>Mean (mm)</th>
<th>Min. (mm)</th>
<th>Max. (mm)</th>
<th>Std. Dev. (mm)</th>
<th>Mean (mm)</th>
<th>Min. (mm)</th>
<th>Max. (mm)</th>
<th>Std. Dev. (mm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>October</td>
<td>65</td>
<td>37</td>
<td>80</td>
<td>12</td>
<td>29</td>
<td>22</td>
<td>32</td>
<td>3</td>
</tr>
<tr>
<td>November</td>
<td>30</td>
<td>9</td>
<td>48</td>
<td>13</td>
<td>6</td>
<td>5</td>
<td>11</td>
<td>2</td>
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<tr>
<td>December</td>
<td>20</td>
<td>9</td>
<td>41</td>
<td>10</td>
<td>1</td>
<td>-0.4</td>
<td>2</td>
<td>0.7</td>
</tr>
<tr>
<td>January</td>
<td>16</td>
<td>0.2</td>
<td>28</td>
<td>10</td>
<td>4</td>
<td>2</td>
<td>5</td>
<td>1</td>
</tr>
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<td>February</td>
<td>13</td>
<td>4</td>
<td>24</td>
<td>6</td>
<td>14</td>
<td>8</td>
<td>20</td>
<td>3</td>
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<tr>
<td>March</td>
<td>12</td>
<td>7</td>
<td>27</td>
<td>6</td>
<td>33</td>
<td>22</td>
<td>42</td>
<td>6</td>
</tr>
<tr>
<td>April</td>
<td>28</td>
<td>16</td>
<td>46</td>
<td>10</td>
<td>72</td>
<td>55</td>
<td>90</td>
<td>11</td>
</tr>
<tr>
<td>May</td>
<td>61</td>
<td>52</td>
<td>79</td>
<td>8</td>
<td>101</td>
<td>90</td>
<td>107</td>
<td>6</td>
</tr>
<tr>
<td>June</td>
<td>101</td>
<td>45</td>
<td>137</td>
<td>27</td>
<td>128</td>
<td>102</td>
<td>153</td>
<td>18</td>
</tr>
<tr>
<td>July</td>
<td>142</td>
<td>122</td>
<td>161</td>
<td>13</td>
<td>148</td>
<td>134</td>
<td>165</td>
<td>11</td>
</tr>
<tr>
<td>August</td>
<td>158</td>
<td>115</td>
<td>194</td>
<td>24</td>
<td>130</td>
<td>109</td>
<td>152</td>
<td>13</td>
</tr>
<tr>
<td>September</td>
<td>108</td>
<td>89</td>
<td>127</td>
<td>14</td>
<td>76</td>
<td>60</td>
<td>95</td>
<td>12</td>
</tr>
<tr>
<td>Average</td>
<td>63</td>
<td>42</td>
<td>83</td>
<td>13</td>
<td>62</td>
<td>51</td>
<td>73</td>
<td>7</td>
</tr>
</tbody>
</table>

Table 3-1 - Mean, minimum, maximum, and standard deviation of monthly evaporation (mm month\textsuperscript{-1}) for Penman and Penman without heat storage.
Figure 3-2 Time series of hourly data averaged for thermal budget periods (TBPs) for the Sooke Reservoir including (A) estimated surface water ($T_w$) and air temperature ($T$, °C); (B) net radiation ($R_n$) and heat storage flux ($Q_{HS}$; MJ m$^{-2}$ d$^{-1}$); (C) wind speed ($U_2$) and adjusted wind speed ($U_a$, m s$^{-1}$); (D) saturated vapour pressure ($e_s$) and vapour pressure deficit ($e$, kPa); and (E) average Penman ($E_P$) and Priestley-Taylor ($E_{PT}$) evaporation (mm d$^{-1}$) for all TBPs.
This high standard deviation, relative to the mean, shows the strong seasonality or intra-annual variability of evaporation rates. Over the PIM, the monthly maximum of 194 mm took place in August 1998, and the minimum of 0.2 mm occurred in January 2004. Note that the standard deviation is highest for the months of June and August at 27 mm month$^{-1}$ and 24 mm month$^{-1}$, respectively. The lowest standard deviations occur in February and March, at 68 and 6 mm month$^{-1}$, respectively suggesting more consistent winter conditions from year to year. The coefficient of variance is largest in January and February (0.65 and 0.53 respectively), and smallest in June and May (0.09 and 0.13). Therefore, the standard deviation is high relative to the mean in January and February and low in June and May.

Annual evaporation totals were computed for October 1$^{st}$ to September 30$^{th}$ water year (Table 3-2). Maximum losses occurred in 1997-1998 at 836 mm and minimum losses of 679 mm took place in 1996-1997. Evaporation for the 2004-2005 water year is not shown because values were not available for the entire year as heat storage estimates were not available for July, August, and September of 2005. Average annual evaporation was 762 mm, with a standard deviation of 53 for the eight available years. However, the coefficient of variation was only 0.07. Hence, the inter-annual variation in evaporation is relatively less than the inter-monthly variation, when compared to the average. The average annual evaporation estimate for the SR compares favourably with the value of 710 mm yr$^{-1}$ for small-lake evaporation estimated by Ferguson (1981) for the region in which the SR is situated.
Figure 3-3 - Monthly evaporation estimates for (A) Penman, (B) Penman without Heat Storage, (C) Priestley-Taylor, and (D) Hamon evaporation methods (mm).
Table 3-2 - Annual totals, mean, minimum, maximum, and standard deviation for Hamon, Priestley-Taylor, Penman, and adjusted Hamon (mm) estimates of evaporation.

<table>
<thead>
<tr>
<th>Water Year</th>
<th>Penman without Heat Storage (mm)</th>
<th>Penman (mm)</th>
<th>Hamon (mm)</th>
<th>Priestley-Taylor (mm)</th>
<th>Adjusted Hamon (mm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>96-97</td>
<td>702</td>
<td>679</td>
<td>599</td>
<td>725</td>
<td>747</td>
</tr>
<tr>
<td>97-98</td>
<td>808</td>
<td>836</td>
<td>653</td>
<td>891</td>
<td>811</td>
</tr>
<tr>
<td>98-99</td>
<td>705</td>
<td>769</td>
<td>593</td>
<td>829</td>
<td>740</td>
</tr>
<tr>
<td>99-00</td>
<td>704</td>
<td>720</td>
<td>599</td>
<td>767</td>
<td>735</td>
</tr>
<tr>
<td>00-01</td>
<td>705</td>
<td>713</td>
<td>580</td>
<td>762</td>
<td>717</td>
</tr>
<tr>
<td>01-02</td>
<td>745</td>
<td>804</td>
<td>588</td>
<td>863</td>
<td>737</td>
</tr>
<tr>
<td>02-03</td>
<td>774</td>
<td>784</td>
<td>631</td>
<td>819</td>
<td>783</td>
</tr>
<tr>
<td>03-04</td>
<td>794</td>
<td>788</td>
<td>635</td>
<td>822</td>
<td>782</td>
</tr>
<tr>
<td><strong>Mean</strong></td>
<td>742</td>
<td>762</td>
<td>610</td>
<td>810</td>
<td>756</td>
</tr>
<tr>
<td><strong>Min</strong></td>
<td>702</td>
<td>680</td>
<td>580</td>
<td>725</td>
<td>717</td>
</tr>
<tr>
<td><strong>Max</strong></td>
<td>808</td>
<td>836</td>
<td>653</td>
<td>891</td>
<td>811</td>
</tr>
<tr>
<td><strong>St. Dev.</strong></td>
<td>44</td>
<td>53</td>
<td>26</td>
<td>55</td>
<td>32</td>
</tr>
</tbody>
</table>

Elevated evaporation rates in 1997-1998 could be the result of the warmer and drier conditions that are attributable to the El Nino event that occurred in combination with a warm PDO (Kiffney et al., 2002; Stahl et al., 2005). The temperature anomaly in 1998 was the largest for 1971-2000 in the SR at 1.4°C above the mean. The 1998-1999 and 1999-2000 water years were associated with moderate and strong La Nina events respectively (MSC, 2005). The PDO was in a cool phase from 1998-2000 (Kiffney et al., 2002). The combination of La Nina and negative PDO has been shown to bring cool, wet conditions to southwest British Columbia (Kiffney et al., 2002; Stahl et al., 2005; Fleming et al., 2007). The below average evaporation amounts for the 1998-1999, 1999-2000, and 2000-2001 water years could be related to strong La Nina events coinciding with the cool phase of the PDO.
Evaporation over the January 2000 to December 2001 period from the SR was estimated to be $10.9 \times 10^6$ m$^3$ by Nowlin et al. (2004), using the Morton (1983) method. Over the same period, evaporation was $7.9 \times 10^6$ m$^3$ based on the Penman method, which is about 27% less than Morton’s. Differences between these two estimates could arise from a number of factors. The Morton (1979) and Penman (1948) methods are significantly different in their approach. Morton’s (1979; 1983) relies solely on monthly values of temperature, humidity and sunshine duration. It takes into account the changes in temperature and humidity of air as it passes from a land environment to a lake environment.

The value provided by Nowlin et al. (2004) is given as a volume (m$^3$). Therefore, differences in the methods used to predict surface area from water level could have contributed to the different estimates produced by the two methods. In addition, heat storage in the lake was approximated with a different method by Nowlin et al. (2004) than in this study and was calculated from different water temperature measurements. Nowlin’s value was converted into mm using the average surface area for January 2000 to December 2001 developed in this study, which resulted in annual evaporation of roughly 1000 mm per year. Evaporation of this magnitude is usually found in areas closer to the equator where net radiation is greater than at mid-latitudes, therefore, it seems unusually high for this region. The absence of monthly or seasonal estimates in Nowlin’s study suggests a lack of confidence in his values for shorter timescales. The methods applied in this study were selected for their ability to replicated seasonal cycles in evaporation. Thus, due to the overestimation of annual values and absence of seasonal data, the results of the Morton approach as applied by Nowlin et al. (2004) are not valuable for this study.

### 4.2 The Priestley-Taylor Estimate

Estimates of evaporation for the TBP are shown in Figure 3-2 – E. The Penman ($E_P$) and Priestley-Taylor ($E_{PT}$) methods show similar timing and magnitudes, although the $E_{PT}$
values are somewhat higher. This difference occurs primarily during the dry season. Figure 3-4 shows a cross-plot of estimates of $E_P$ and $E_{PT}$ for the TBPs.

![Figure 3-4 - Cross-plot of average Priestley-Taylor ($E_{PT}$) and Penman ($E_P$) evaporation (mm d$^{-1}$) estimates for thermal budget periods (TBPs).](image)

The relationship between the two was strong with a Pearson’s $r = 0.99$. A linear regression between the two estimates yielded $E_{PT} = 1.08 E_P - 0.08$, which shows that the $E_P$ estimates are more conservative (by roughly 8%) than those from the $E_{PT}$. Other studies have found that $E_{PT}$ estimates are often slightly larger than $E_P$ estimates (Mosner and Aulenback, 2003; Winter et al., 2005).

While $E_P$ and $E_{PT}$ both include net radiation ($R_n$), heat storage ($Q_{HS}$), the slope of the saturation vapour pressure curve ($s$), the psychrometric coefficient ($\gamma$), and the latent heat of vaporization ($\lambda$) in their equations, $E_{PT}$ does not include the wind speed function ($f_u$) or the vapour pressure deficit ($D$). $E_{PT}$ instead includes $\nu$ with a set value of 1.26. According to this value of $\nu$, the aerodynamic influence on the evaporation process is 26% of the available-energy influence (Rosenberry et al., 2007). This average value is representative of the sites
and time period for which it was developed. For example, when the aerodynamic influence on evaporation is less than this empirically determined value, the Priestley-Taylor $\nu$ of 1.26 will cause evaporation to be overestimated.

In many studies, $\nu$ has been estimated empirically, specifically for the site at which it was applied. Other studies have found $\nu$ to be as low as 1.035 (Slouch et al., 1996). Rosenberry et al. (2007) found that a $\nu$ of 1.235 would be required to eliminate the bias between the Priestley-Taylor and Bowen Ratio Energy Budget method in their study on Mirror Lake, New Hampshire, USA. Estimating $\nu$ was not the focus of this work. However, to establish the cause for the discrepancies between $E_p$ and $E_{PT}$, alternative values of $\nu$ were explored. Interestingly, we might suggest a $\nu$ of 1.17 to close the gap between $E_p$ and $E_{PT}$ in this study.

Another possible cause for the discrepancy between $E_p$ and $E_{PT}$ is the $au$ value in the $fu$ term. As described in Section 3.2 this value was set at 0.5 instead of 1.0 as originally suggested by Penman (1948). Work by Penman (1956), Brutsaert (1982), and Cohen et al. (2002) suggested that 0.5 was more realistic over open water. Large water bodies have increased aerodynamic resistance that prevents the diffusion of water vapour, which is inversely proportional to $fu$. Hence, a lower $au$ is more suitable for large water bodies like the SR. The influence of $au$ on the difference between $E_p$ and $E_{PT}$ was tested by changing $au$ from 0.5 to 1.0 while maintain a $\nu$ of 1.26 in $E_{PT}$. With this adjustment, the regression line between the two estimates became $E_{PT} = 1.03E_p - 0.08 \ (R^2 = 0.98)$, decreasing the gap between the two estimates by 5%. This gives further evidence that the $\nu$ of 1.26 in $E_{PT}$ likely contributes to the overestimation of evaporation from the SR.

Differences between $E_{PT}$ and $E_p$ were larger in some months than others. In June, September, November, and December $E_{PT}$ was greater than $E_p$ and in February, $E_p$ was greater than $E_{PT}$. This suggests that $\nu$ value of 1.26 in $E_{PT}$ overestimated the aerodynamic
influence on evaporation in June, September, November, and December and underestimated it in February. In months where $E_{PT}$ was greater than $E_P$, wind speed ($U_2$) was likely lower and relative humidity ($RH$) was likely higher than they were in the study sites where the $\nu$ value of 1.26 for $E_{PT}$ was established. Relative humidity, $RH$, is important because it is a primary determinant of $D$ (Eq. 2). In February, it is possible that $U_2$ is greater or $RH$ is less in the SR than they were in the sites where $E_{PT}$ was developed, which would cause more evaporation to be estimated with the $E_P$ method. The northern Mediterranean climate of the SR may have stronger seasonal contrasts in $U_2$ and $RH$ than the sites where 1.26 was established as the value for $\nu$ in $E_{PT}$.

Considering that the $RH$ was not measured at the recommended 2 m above the water surface but instead below the dam, where there is likely to be less moisture, it was expected that the $RH$ might be somewhat underestimated during the summer months for the SR. The dam itself could be creating a boundary over which moist air does not pass easily from the lake surface to the surrounding land. However, increasing $RH$ would have decreased $E_P$ and increased the difference between it and $E_{PT}$. To explore the possible influence of a decrease in $RH$, which would be expected to increase evaporation to match $E_P$ to $E_{PT}$, $RH$ was decreased by 10% on those days when $RH$ was < 80% and the difference between $E_P$ and $E_{PT}$ reduced by only 3%.

Based on these results, it appears that the major difference between the two estimates is attributable to $\nu$. A preliminary adjustment of $\nu$ and then of $a_u$ resulted in the $E_P$ and $E_{PT}$ estimates replicating each other more closely. Hence, it has been established based on comparing $E_P$ to $E_{PT}$ that the measurement of $U_2$ and $RH$ at heights other than the recommended 2 m above the lake surface has not significantly diminished the ability of the $E_P$ method to provide representative values of evaporation. This could be because the wind-speed measurements were adjusted before they were used in $E_P$. Findings from previous
studies have shown that a modified form of the Penman method can provide reasonable
estimates using wind speed and relative humidity values that were even taken from the shore
of the lake (Winter et al., 2005; Valiantzas, 2006).

4.3 Heat-Storage Flux

Heat-storage flux, $Q_{HS}$, is included in both the Penman and Priestley-Taylor methods.
$Q_{HS}$ typically ranged over the year from -8.6 MJ d$^{-1}$ to 8.6 MJ d$^{-1}$ with some exceptional
events, such as the 12.3 MJ d$^{-1}$ reached in May 2005 (Figure 3-2 – B). Positive $Q_{HS}$ values
were at a maximum in May or June and decreased thereafter, eventually switching to negative
in July or August. Negative $Q_{HS}$ reached a minimum in August or October of most years. As a
result, the SR gained most of its heat between late-February/March and late-July and lost
most of its heat between late-August/October and February.

The seasonal variation and magnitude of $Q_{HS}$ for the SR were comparable to those
found in other studies (Derecki, 1975; USGS, 1954; Winter et al., 1995; Lenters et al., 2003).
One notable difference between $Q_{HS}$ estimates for the SR and those for other studies was that
many of the other studies did not estimate $Q_{HS}$ for the November to March period, as the
lakes were often ice covered. Permanent ice cover was not recorded in the SR during the
PIM. During the November to March period, $Q_{HS}$ was predominately negative and
evaporation was often occurring in spite of low net radiation. Heat that had been stored in the
lake during the previous high-radiation months was released during these cool, low net-
radiation periods, thereby causing evaporation. The importance of the release of heat storage
in driving evaporation has been demonstrated in other studies where fluxes were measured
more directly (e.g. Blanken et al. (2000a) which employed the eddy co-variance method).

$Q_{HS}$ is controlled by a multitude of factors such as net radiation, wind speed, and
synoptic patterns (Blanken et al., 2000a; Blanken et al., 2000b). Quiescent periods contribute
to rising surface water temperatures when there is high net radiation. Increased wind
following these periods can effectively mix warmer waters downward (Blanken et al., 2000a). The influence of wind speed on heat storage was investigated in the SR where daily $Q_{\text{HS}}$ was available (Figure 3-5).

\begin{figure}[h]
\centering
\includegraphics[width=\textwidth]{figure3-5.png}
\caption{Heat storage (Joules) calculated from a thermoster string at site SOL_01 and adjusted wind speed ($U_a$; m s$^{-1}$) for August 2003 to September 2004.}
\end{figure}

Generally, from mid-October to mid-March, wind speeds were high and $Q_{\text{HS}}$ was low. After March, wind speeds subsided, allowing net radiation to penetrate the SR surface and increase the heat storage. Without more precise measurements (i.e., every 15 minutes), the exact relationship between wind speed and $Q_{\text{HS}}$ cannot be analyzed for the SR. However, it is important to note that climate change may include changes to predominant synoptic patterns in the area that effect wind speed. These adjustments in wind speed could influence heat-storage patterns and in turn, modify evaporation.

Other factors influencing $Q_{\text{HS}}$ include heat advected by a water source such as precipitation, streamflow, or outflows. For the large part, these have been assumed to balance to zero. However, initial estimates of evaporation using the Penman method showed a few TBPs that had large negative values of evaporation, or condensation. Evaporation estimates probably resulted in condensation in these TBPs because the positive $Q_{\text{HS}}$ values were larger.
than the positive net radiation values, and when subtracted from the net radiation values, negative evaporation resulted in Eq. (1). These high $Q_{HS}$ events coincided with above normal precipitation coupled with warm air temperatures, suggesting that the precipitation was warm and that it had advected a large amount of heat into the SR.

One of these warm, high precipitation events took place from October 14 to October 21, 2003. During this eight-day period, 481 mm of rainfall occurred, which was 28% of the total 1725 mm of rainfall for that year. Two unusually large events took place in this period: (1) 243 mm on October 16 2003; and (2) 115 mm on October 20 2003. The average air temperature for this period was 11.7°C, with a low of 8.4°C on October 14, 2003 and a high of 14°C on October 20, 2003. This period was warmer and wetter than average for this time of year, based on records from the PIM, and was likely caused by a “pineapple express”. A “pineapple express” is a Pacific Ocean subtropical jet stream that brings warm moist air from Hawaii to the west coast of North America and has a strong influence over the southwest coast of BC (Stahl et al., 2005; Moore et al., forthcoming 2007). Therefore, anomalous $Q_{HS}$ values were replaced with the average $Q_{HS}$ from the month before and after the anomalous value. There were four cases where records were infilled TBP 784, 791, 2552 and 3013, November 12 to 18, 1998, November 18 to 25, 1998, October 14 to 21, 2003, and January 19 to 24, 2005, respectively.

Other noticeable features of $Q_{HS}$ are the changes to the timing and magnitude that seemed to occur near the beginning of 2003. Two things happened around this time that could have led to these changes: (1) in December 2002, dam construction was complete and the SR volume was increased by 78%; and (2) after April 2003, water temperature profiles were taken more regularly (every eight days on average, rather than every 13 on average). Changing the volume of the dam could have influenced how water in the SR was heated. Larger water volumes would have required more net radiation to achieve the same amount of warming and would have taken longer to heat. Hence, the increase in heat storage would have
been delayed. Once the volume of the dam was increased, the rate and timing of withdrawal and spill from the SR was adjusted. Namely, before the dam was raised, large volumes of water were spilled from the SR in the rainy months of January, February and March, but after the building of the dam, it was no longer necessary to spill. When this cool, winter water was no longer spilled, decreases in $Q_{HS}$ could have resulted. Increasing the frequency of water temperature profiles could have led to the capture of more water-temperature variability, which resulted in increased $Q_{HS}$ variability. Since both of these changes took place at approximately the same time, the influence of one versus the other on $Q_{HS}$ cannot be distinguished.

The influence of $Q_{HS}$ on evaporation estimates was tested with the Penman method, $E_P$, because $E_P$ was considered more robust than the Priestley-Taylor method for estimating open-water evaporation (Winter et. al, 1995). Figure 3-6 shows $E_P$ estimated with and without the heat-storage flux term, $Q_{HS}$. When not included, evaporation follows the timing and magnitude of net radiation, peaking in July and reaching its lowest rates in January. Including $Q_{HS}$ in $E_P$ shifts the evaporation peak to August and increases the evaporation in December/January by driving evaporation with heat released from the SR. Rates can be up to three orders of magnitude greater during January when $Q_{HS}$ is included in the calculation.

![Figure 3-6 - Monthly evaporation (mm) estimated with the Penman ($E_P$) method with and without heat storage ($Q_{HS}$).](image_url)
The results of a t-test (Table 3-3) comparing monthly $E_P$ estimated with and without $Q_{HS}$ showed that the difference in means between the two were significant at the 95% confidence level for 10 of the 12 months. It should be noted that the sample size for all t-tests in this study were small, given data were only available for nine years. Including $Q_{HS}$ in Eq. (1) resulted in higher evaporation rates from August to January (inclusive), differing significantly from $E_P$ estimates without $Q_{HS}$. From February through to July (inclusive), $E_P$ rates were less when $Q_{HS}$ was included. All but two of the months, February and July, had significant differences. $E_P$ was as much as 36 mm (October) greater and as much as 43 mm (April) less, when $Q_{HS}$ was included.

<table>
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<th>Month</th>
<th>$t$-statistic</th>
<th>p-value</th>
<th>Difference value (mm)</th>
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<th>upper bound (mm)</th>
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Table 3-3 - Difference in means (mm) for Penman minus Penman without heat storage. Significant differences are marked with an asterisk (*).

$Q_{HS}$ is an important component of the energy budget for evaporation in the SR because of the way it modifies the timing and magnitude of evaporation, especially at monthly and seasonal timescales. These results could have strong implications for managing water supplies in the dry season, especially during August, September, and October, when the loss of supply through evaporation could be underestimated by techniques that do not include $Q_{HS}$. Because of its influence on evaporation, heat storage should be accounted for in the
reconstruction of historical evaporation rates and considered in any future estimates of evaporation.

4.4 The Hamon and Monthly Adjustment Factors

Evaporation was estimated with the Hamon method \((E_H)\) on an annual and monthly basis and compared to \(E_P\). Annual \(E_H\) ranged from a low of 580 mm in the 2000-2001 water year to a high of 653 mm in 1997-1998 and was 610 mm per year on average (Table 3-2). On average annual \(E_H\) was only 80% of \(E_P\) with differences as high as 215 mm and as low as 80 mm annually. Average differences were 152 mm. Monthly \(E_H\) ranged from a minimum of 14 mm in December 1996 to a maximum of 109 mm in July 1998, and was 51± 28 mm on average with a coefficient of variation of 0.54 (Table 3-4; Figure 3-3 – D).

<table>
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<th>Max. (mm)</th>
<th>Std. Dev. (mm)</th>
<th>Mean (mm)</th>
<th>Min. (mm)</th>
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<tr>
<td>Average</td>
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<td>14</td>
<td>51</td>
<td>46</td>
<td>56</td>
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Table 3-4 - Mean, minimum, maximum, and standard deviation of monthly evaporation for Priestley-Taylor and Hamon (mm month\(^{-1}\)).

Compared to the minimum monthly \(E_P\) of 0.2 mm in January 2004 and the maximum of 194 mm in August 1998 (when \(Q_{HS}\) was included), the range in \(E_H\) was less than that for \(E_P\). Minimum and maximum values for these two methods also occurred in different months. Average monthly values for \(E_P\) were 61± 51 mm with a coefficient of variation of 0.84, which is 36% greater than the coefficient of variation for \(E_H\) of 0.54.
To further assess the monthly differences, a t-test of the difference in means between the monthly $E_P$ and $E_H$ was conducted. Results revealed that differences were significant ($p<0.05$) for nine of the 12 months (Table 5). November, December, and January were the only months that were not significantly different. From January to May $E_H$ was > than $E_P$ and from June to December $E_H$ was < than $E_P$. Therefore, $E_H$ underestimated evaporation during the dry season and overestimated evaporation during the wet season.

<table>
<thead>
<tr>
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<th>p-value</th>
<th>difference value (mm)</th>
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<th>upper bound (mm)</th>
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</tr>
<tr>
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<tr>
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<td>-8</td>
<td>9</td>
<td>0.4</td>
</tr>
</tbody>
</table>

Table 3-5 - Difference in means (mm) for Penman minus Hamon estimates of evaporation. Significant differences are marked with an asterisk (*).

Monthly $E_H$ estimates did not match the timing, magnitude, and variance of $E_P$. $Q_{HS}$ was shown to influence the timing and magnitude of $E_P$ (see Section 4.3) and to increase the coefficient of variation of $E_P$. Figure 3-7 compares $E_P$, with and without $Q_{HS}$ included, to $E_H$. $E_H$ followed the timing of $E_P$ when $Q_{HS}$ was included more closely. For example, both peaked in July. From this we can see that a large influence on the difference between the seasonality of the monthly $E_H$ and $E_P$ estimates likely arises from $Q_{HS}$ not being accounted for in $E_H$ as it has been in $E_P$. Hamon (1961) compared $E_P$ and $E_H$ on a monthly timescale for evapo-transpiration and showed values to be similar. However, evapo-transpiration is not influenced by heat-storage effects as much open-water evaporation, and $E_H$ was of a slightly different
form in that older version, where $L$ was squared, making the function less linear than Eq. (27).

Along with the difference between $E_H$ and $E_P$ on a monthly or seasonal timescale, annually $E_H$ was consistently less than $E_P$ by 152 mm on average. This difference is outside the range of -76.2 to 76.2 mm found by Hamon (1961) when he compared these two methods for lake evaporation. In the 1961 paper, testing of the $E_H$ technique was carried out in the eastern and southern US, areas that have different hydroclimatic settings than the SR. This could account somewhat for the larger difference found in the SR. $Q_{HS}$ could be partially responsible for these differences because, as shown in Table 2, annual $E_P$ estimated with $Q_{HS}$ included was almost always slightly greater than without. Thereby, $Q_{HS}$ primarily influenced the seasonality, but also slightly increased the annual values. Comparison of Eqs. (5) and (27) shows that in addition to $Q_{HS}$ not being included in $E_H$; $R_n$, $r_a$ and $D$ are represented in $E_H$ by surrogates, namely $T_a$, $L$, $e_a^*$, and a fixed coefficient of 2.98 a value. Dingman (1994) reformatted Hamon’s work to derive these parameters and values. Therefore, measurements of $f_a$, $RH$, and $R_n$ are reflected in the $E_H$ estimate. If the relationship between these variables
and their surrogates in Eq. (27) are not the same in the SR as they were where Eq. (27) was developed, different estimates might result.

Given the nine years of overlapping estimates of $E_P$ and $E_H$ available over the PIM, a method was devised to adjust $E_H$ to emulate $E_P$ more closely. Since the difference between the two estimates is likely driven by the inclusion of $Q_{HS}$ in $E_P$ the modified form, which accounts for heat storage is denoted as $E_{HSC}$. The average monthly ratio between the two methods was derived by calculating the $E_P$ to $E_H$ ratio on a monthly basis as follows:

$$E_{HSC} = E_H \cdot \psi_m \text{ (mm month}^{-1}\text{)}$$  \hspace{1cm} (28)

where $E_{HSC}$ is the adjusted evaporation estimate and $\psi_m$ is the average percentage ratio between the $E_P$ and $E_H$ estimates over the PIM for a given month, $m$.

The monthly values for the $\psi_m$ and the root mean squared error (RMSE) between $E_P$ and $E_{HSC}$ for each month are provided in Figure 3-8. The RMSE values are highest for June (256.26) and lowest for February (18.82). The large RMSE for June, when compared to other months, likely results from high inter-annual variability of $E_P$ in June (Table 3-1 and Figure 3-9). Other months, such as August, also have high inter-annual variability in $E_P$ (Table 3-1) and similarly have large RMSE between $E_P$ and $E_{HSC}$. This level of inter-annual variability in $E_P$ might be related to inter-annual variability in the variables that are used to estimate $E_P$. For example, $R_n$ has the highest standard deviation in the month of June (Table 3-6). Therefore, this relationship does not work as well in months where $E_P$ has large inter-annual variability. Conversely, in months such as February, where the standard deviation of $E_P$ is low, RMSE is relatively low (Table 3-1; Figure 3-8; Figure 3-9). Based on this analysis, this conversion approach is not strong in the months of June and August, but is satisfactory for the remaining months.
Figure 3-8 - Box and whisker plots of monthly Hamon and Penman evaporation estimates (mm). Whisker show minimum (bottom) and maximum (top) values, dark black line shows median, extent of box shows 25th and 75th percentiles, and circles show outliers.

Figure 3-9 - Average percentage difference (%) of Penman versus Hamon evaporation estimates by month and the RMSE for each month of the resulting adjusted-Hamon estimates when compared to the Penman estimates.
Changes to heat-storage patterns seen after December 2002 added complexity to this situation. As discussed in Section 4.3, \( Q_{HS} \) patterns changed slightly around this time and therefore could have influenced \( E_P \) and, in turn, the relationship of \( E_P \) to \( E_H \) and the resulting estimates of \( E_{HSC} \). It cannot be determined whether these changes were related to the increase in capacity or an increase in the frequency of water temperature measurement, but it should be noted that the dam was modified twice in the past and it is possible that these changes also affected \( Q_{HS} \). Because of this, the \( \psi_m \) may not provide reliable estimates of \( E_{HSC} \) for each phase of the dam back to 1916.

<table>
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<tr>
<th>Month</th>
<th>( Q_{HS} ) (MJ m(^{-2}) day(^{-1}))</th>
<th>( Q_{HS} ) (MJ m(^{-2}) day(^{-1}))</th>
<th>( T_{min} ) ((^{\circ})C)</th>
<th>( T_{min} ) ((^{\circ})C)</th>
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<th>( RH ) (%)</th>
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<th>( U_h ) (m s(^{-1}))</th>
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<td>2.6</td>
<td>0.2</td>
<td>1.8</td>
<td>0.1</td>
<td>13.0</td>
<td>0.9</td>
</tr>
<tr>
<td>Sep</td>
<td>82.6</td>
<td>5.1</td>
<td>2.3</td>
<td>0.2</td>
<td>1.6</td>
<td>0.1</td>
<td>8.3</td>
<td>1.0</td>
</tr>
<tr>
<td>Oct</td>
<td>92.4</td>
<td>1.7</td>
<td>2.4</td>
<td>0.2</td>
<td>1.6</td>
<td>0.1</td>
<td>3.4</td>
<td>0.3</td>
</tr>
<tr>
<td>Nov</td>
<td>96.9</td>
<td>1.9</td>
<td>2.5</td>
<td>0.3</td>
<td>1.7</td>
<td>0.2</td>
<td>0.8</td>
<td>0.2</td>
</tr>
<tr>
<td>Dec</td>
<td>97.3</td>
<td>1.2</td>
<td>2.9</td>
<td>0.4</td>
<td>2.0</td>
<td>0.3</td>
<td>-0.1</td>
<td>0.0</td>
</tr>
</tbody>
</table>

Table 3-6 - Mean and standard deviation of monthly heat storage (\( Q_{HS} \); MJ m\(^{-2}\) d\(^{-1}\)), minimum air temperature (\( T_{min} \); \(^{\circ}\)C), maximum air temperature (\( T_{max} \); \(^{\circ}\)C), mean air temperature (\( T \); \(^{\circ}\)C), relative humidity (\( RH \); %), wind speed (\( U_h \); m s\(^{-1}\)), adjusted windspeed (\( U_a \); m s\(^{-1}\)), and net radiation (\( R_n \); MJ m\(^{-2}\) d\(^{-1}\)).
Monthly values of $E_H$, $E_P$ and $E_{HSC}$ for the PIM are compared in Figure 3-10. Inspection of the plot shows that $E_{HSC}$ is a definite improvement over $E_H$. The timing and magnitude of $E_P$ is better replicated with $E_{HSC}$ than it was with $E_H$. For many of the months, $E_P$ and $E_{HSC}$ are quite close (RMSE < 41; Figure 3-9), such as February to May, and July. June, August, October and November do not match as closely, and values are not consistently over- or under-estimated. Overall, this adjustment technique provides monthly evaporation estimates that are an improvement over $E_H$ and as a result, this technique may be valuable for estimating open-water evaporation where climatic data is limited, such as reconstructions of past climate.

![Figure 3-10 - Monthly evaporation estimated via the Penman ($E_P$), Hamon ($E_H$), and adjusted Hamon ($E_{HSC}$) methods (mm).](image)

5 CONCLUSIONS AND RECOMMENDATIONS

Estimation of evaporation from the SR indicates that there is a strong seasonal pattern in the evaporation rates. Strong seasonal effects produce large evaporation volumes over the dry season (April-September) and low volumes in the wet season (October-March). Water leaving the SR via evaporation during the dry season could therefore play an important role in the water balance by potentially outweighing inputs from precipitation and streamflow. The standard deviation of evaporation rates is largest in the months of June and August. Evaporation rates for these months could be strongly over- or under-estimated if they are
predicted based on annual averages. Therefore, monthly or seasonal evaporation is an important component of the water balance and should be considered in management strategies.

Heat storage in the SR had a significant impact on the timing and magnitude of evaporation by delaying maximum evaporation by roughly one month and of extending evaporation into the winter. Therefore, it is important to account for heat-storage fluxes to estimate evaporation rates on a daily or monthly basis. Comparison of $E_H$ to $E_P$ showed that $E_H$ evaporation estimates, which were based on air temperature and day-length, underestimated evaporation by an average of 151.8 mm annually and did not match the seasonal pattern displayed by $E_P$. Annual $E_H$ values were only 80% of $E_P$ on average. Evaporation was overestimated during the wet season and underestimated during the dry season by $E_H$. The discrepancies in the monthly timing and magnitude were partly due to $Q_{HS}$ not being accounted for by $E_H$. Factors such as $R_n$, $r_a$ and $D$ being represented in $E_H$ by surrogates, namely $T_a$, $L$, $e_a^*$ and a fixed coefficient of 2.98 (reformed by Dingman based on values developed empirically by Hamon), likely contributed to the discrepancies between $E_H$ and $E_P$. Considering the importance of evaporation to the water balance during the dry season, underestimates could strongly reduce the probability of predicting low-water events resulting from water loss via evaporation. Therefore, it is recommended that $E_H$ not be used to estimate open-water evaporation from the SR in its original form. This is to be expected as it was developed originally to estimate potential evaporation from well-watered landscapes and not for estimating evaporation from large, deep reservoirs, like the SR.

The $E_{HSC}$ values emulated $E_P$ values successfully in months where $E_P$ had lower inter-annual variability, such as February to May, and July. June, August, October and November were no as well replicated because of the high inter-annual variability in $E_P$ for these months, which probably resulted from variability in $R_n$ and $Q_{HS}$ for year to year for these months. Changes to the timing of water-temperature measurement and to the capacity of the dam had
an influence on estimated $Q_{HS}$. The adjusted Hamon method can be used to hind-cast evaporation back to 1916, and to project evaporation amounts for the future. It should be considered, however, that different dam configurations could influence heat-storage behaviour. Future work should test a multiple-linear regression equation that could explain the relationship between $T_a$ and $Q_{HS}$, $R_n$, and $D$ to develop a more robust $E_{HSC}$ estimate that would provide reasonable values under varying reservoir capacities.

This is one of few in-depth studies of lake evaporation in North America, which was conducted for an extended period (almost 10 years) on a lake that has open water for the entire year. Many other studies have reported evaporation during the months of May through to November only. Evaporation rates from the SR for the months of May to November were comparable to monthly mean evaporation rates for lakes in the United States: Sparkling Lake (Lenters et al., 2005); Perch Lake (Robertson and Barry, 1985); Williams Lake (Sturrock et al., 1992); and Mirror Lake (Winter et al., 2003). Evaporation in the SR during fall and early winter can be substantial because heat stored in the lake drives evaporation. In comparison with two studies conducted at more southern latitudes with mild winters (Westenburg et al., 2006; Riley et al., 2006), evaporation in the SR was found to be less during the shoulder seasons (April-May and September-October). This is likely because of enhanced evaporation in these areas as a result of more net radiation.

Within the short PIM period, it appears that high evaporation rates in the SR occurred during the warm, dry years in the record, such as those during the El Nino phase of ENSO. Lower evaporation rates occurred during the cool, wet La Nina events. The climatological trend over the past century has been towards increasing temperatures. The IPCC (2007) reported that 11 of the 12 years from 1995 to 2006 rank among the twelve warmest years in the instrumental record of global surface temperature (since 1850). If greenhouse gas emissions continue at current rates or increase, further warming will be very likely (IPCC, 2007).
Projections for the next two decades are for a warming of about 0.2°C per decade globally, based on a range of emission scenarios. Even if the concentrations of all greenhouse gases and aerosols are kept constant at year 2000 levels, a further warming of about 0.1°C per decade would be expected (IPCC, 2007). Rates of change of temperature and precipitation in southern BC and the Pacific Northwest are greater than global averages (Mote, 2003a; Mote, 2003b; Zhang et al., 2000). Considering this, evaporation is likely to draw strongly on water reserves in the future and approaches to managing systems under these conditions should be considered.

Future work to improve the estimation and possible prediction of evaporation should include improving the understanding of the relationship of evaporation rates to synoptic weather events and to modes of climate variability. Measurements of the temperatures of inflows and outflows are available and could be used to improve estimates. They would serve as a check by monitoring the transfer of energy through the processes of advection. Estimates of evaporation could also be improved by augmenting the current meteorological measurements made in the SRC. Heat storage estimates could be improved by taking temperature profiles on a more frequent schedule, such as every seven days. While the adjustment of the wind speed using a profile method was adequate for the purposes of this study, future work might benefit from wind speed and relative humidity measurements being made on the surface of the SR at the prescribed 2 m height above the lake surface. Placing these stations on the water surface would also reduce the influence of the dam on the gauging of the meteorological variables. Future estimates of evaporation should be made with the Penman method because it is the most seasonally accurate.
REFERENCES


Capital Regional District (CRD) 2006, CRD Water Services - Sooke Reservoir Project Description http://www.crd.bc.ca/water/engineering/sookereservoir/description.htm


CHAPTER 4: SEASONALITY OF THE WATER BALANCE OF THE SOOKE RESERVOIR, BC, CANADA

ABSTRACT

Precipitation, runoff, consumption, spill and evaporation at the Sooke Reservoir are investigated in order to close the water balance and understand its seasonality over the nine-year period of intensive monitoring (October 1996 to September 2005). A worst-case scenario of drought is constructed by combining the wet season with the least precipitation during this period and the dry season with the highest evaporation to assess the time it takes for water storage to fall below accessible levels. Since total inflows were not previously known, the Hydrologiska Byrans Vattenbalansavdelning hydrologic model is tested for estimating total inflows against the contributing area approach. Evaporation is provided from a companion study using the Penman method, while precipitation, consumption, spill, and storage are measured by the Capital Regional District, which manages the supply.

The Sooke Reservoir is a rain-dominated system where the majority of its supply arrives over the winter period via rain and run-off. There are two distinct seasons, the wet season (October to March) and the dry season (April to September). During the dry season, evaporation accounts for 9% of the water loss from the system, outweighing inputs from precipitation for eight of the nine years in the period when intense monitoring took place. In the worst-case drought scenario, the storage reaches inaccessible levels after three years of persistent low-water wet seasons and high-evaporation dry seasons, as derived directly from cases in the study period. Monthly values of the water balance components defined in this study provide a foundation for studying climate impacts on this water supply in the future.
1 INTRODUCTION

The Sooke Reservoir (SR) is a municipal reservoir that supplies the Greater Victoria Area. The southern portion of Vancouver Island, which includes the SR, has been described as one of the most hydro-climatically complex in British Columbia, primarily due to its close proximity to the Pacific Ocean and mountainous terrain (Wade et al., 2001; Fleming et al., 2007; Werner et al., 2007a). The seasonality of this region is considerable, with distinct wet (October to March) and dry periods (April to September), which classifies it as a northern Mediterranean regime ($Cs$) by Koeppen’s (1958) system.

The SR has been in operation for close to a century and has adequately supplied the residents of the Greater Victoria Area for most of this time. However, within the last decade, a number of events have occurred when the ability of the Capital Regional District (CRD) to manage the supply has been challenged. In 1998 and 2001, low water-availability developed from a combination of changes in climate and increases in demand from a growing population. These two factors are likely to increase pressure on supply in the future.

In the past, limitations of the SR supply have been addressed by (1) demand-side management practices; and (2) identifying a possible future water source. With demand-side management, consumers control water usage through water conservation methods and incentives, which are partially driven by the economy (Wolff and Glieck, 2002). These types of strategies, which reduce water use by encouraging behavioural changes in the consumers, will help to alleviate the problem of water limitations, but may not be a long-term solution. The potential future water source for the SR has been identified as the nearby Leech River. However, it has less than ideal water quality and a limited capacity to augment supply during low-flow periods (Kolisnek, 2006). Hence, further study of the water balance of the SR would aid in more efficient management of this supply to prevent the need to rely on the Leech River as the sole solution.
Climate change and variability are likely to have strong implications for reservoir operation because of resulting changes to the seasonal patterns of precipitation, air temperature and streamflow (Hamlet and Lettenmaier, 1999). Understanding these shifts helps managers of a water supply to optimize their operation to meet demand (Regonda et al., 2005). However, investigation of the influence of climate change and variability require long-term records (>30 years). In the SR, longer-term records of precipitation and temperature exist, but the collection of other hydro-climatically relevant data is limited to the near term.

To understand how a reservoir will react to climate change and variability, it is necessary to know each component separately, as the response of the different components of its water balance may differ and magnify the effect. For example, the response of streamflow to climate variability sometimes lags or magnifies changes in air temperature or precipitation (Kiffney et al., 1999; Fleming et al., 2007). Knowledge of the near-term water balance could be used to relate storage in the SR to the longer-term records that are available, such as temperature and precipitation.

The capacity of the SR cannot be increased any further and climate change and population growth are increasing pressure on the supply. Furthermore, current adaptation measures are not satisfactory and challenging to devise without fully understanding the water balance. These factors indicate that the SR should be monitored carefully so that it can be managed within its current limits. The primary objective of this study is to understand the water balance components of the SR and their seasonality. This objective is pursued by evaluating each component individually on a monthly and seasonal basis. The water balance of the SR has been defined as (Werner, 2007a):

\[ I_s + I_G + P - E - O_{s1} - O_{s2} - O_{s3} - O_G \pm \Delta \varepsilon = \Delta S \] (m³ month⁻¹) (1)
where $I_s$ is surface inflow, $I_G$ is groundwater inflow, $P$ is precipitation onto the SR, $E$ is evaporation from the SR, $O_{S1}$ is consumption, $O_{S2}$ is spill, $O_{S3}$ is fisheries release, $O_G$ is groundwater outflow, $\Delta S$ is change in storage, and $\varepsilon$ is the error accumulated from each term (all presented in m$^3$ month$^{-1}$). Records of air temperature, precipitation, $O_{S1}$, $O_{S2}$, and water level have been taken since 1916. However, data pertaining to the majority of the components of the water balance have been gathered most intensively and with greater precision from October 1996 to September 2005. Therefore, this will be the period of study for this paper and will be referred to as the “period of intensive monitoring,” or PIM.

The components of the water balance are known to varying degrees of magnitude and accuracy. Many are measured directly by the CRD, such as $O_{S1}$, $O_{S2}$, $O_{S3}$ and $\Delta S$. $E$ was investigated thoroughly by Werner (2007b). $P$ was studied by Neimann (1981) and Fairbairn (2003) as described in Werner (2007a). $I_G$ and $O_G$ are unknown. Inflows are gauged for 42% of the SR catchment (SRC), but the remaining 58% has not been estimated independently. Estimating total inflows to the SR is therefore a sub-objective of this study.

Although extremes are the focus of much work on climate change and its effect on water supply, they have not been evaluated for the SR with the exception of a precipitation study by Dore et al. (2005). The amount of storage available for use in the SR results from multiple factors. Hence, the effect of extreme weather and climate on storage in the SR can be measured only with a solid understanding of the interaction of the water balance components. If one diverges in one direction, another may serve to mitigate or enhance the effect. Thus, the second objective of this work is to test the resilience of the SR system to extremes. This is done by creating a worst-case drought scenario by combining the lowest precipitation wet season with the highest evaporation dry season. By this method, drawdown in the SR during times of persistent extreme warm and dry conditions is simulated.
2 STUDY AREA

The SR drainage basin (Figure 4-1) is located at 48° 30' 50" N latitude and 123° 42' 1" W longitude. Covering an area of 70.1 km² (including the 7.1 km² SR), it represents about 91% of the Greater Victoria Water District’s (GVWD) water supply (CRD, 2006). Water is diverted from the Council Creek watershed to supplement the Sooke Reservoir Catchment (SRC) supply, which adds 10 km² of drainage area when in use (Green and Gillie, 1994).

The SR has a northern Mediterranean climate, more broadly classified as a Cs climatic type with distinct wet and dry seasons, and mild winters (Koeppen and De Long, 1958). These conditions arise from patterns of low pressure (cyclonic) and high-pressure (anticyclonic) systems that alternate within the year (Moore et al., forthcoming 2007). The close proximity to the Pacific Ocean and mid-latitude location makes this region susceptible to changes in storm tracks and onshore winds brought on by westerlies (low-pressure systems; Bryson and Hare, 1974). High pressure limits precipitation and promotes warmer temperatures in the summer (Chilton, 2000). Consequently, most of the precipitation falls between October and April in this region, and the remaining months are relatively dry.

The hydrologic response of the SR watershed is reflective of the above noted hydroclimatic characteristics. In general, it can be classified as a predominately pluvial regime where early winter (November or December) is typified by a large hydrograph peak generated from winter rains (sometimes augmented by snowmelt at higher elevations) and is followed by consistent rainfall runoff over the remainder of the winter. This wet season often spans from November to March or April, and is common to mid-altitude watersheds in the coast mountain region (Whitfield et al., 2003; Fleming et al., 2007). If there is some winter accumulation of snow, it does not persist long enough to augment flows in the spring or summer. The dry season begins in April as rainfall tapers off and air temperature begins to increase. Dry conditions prevail August through September characterized by an extended
period of minimal rainfall concomitant with higher temperatures. In late spring through to early fall, inflows diminish while water consumption increases, causing the SR water level to decline until the rains arrive again in late fall (CRD 2005).

The geometric configuration of the SR has changed from a historical series of increases in dam-retention heights, the second and last occurring in 2002 when it was raised 6 m to a height of 186.75 metres above sea level (masl) increasing the holding capacity by 78% (92.7 x 10^6 m³ from 52.0 x 10^6 m³). A secondary reservoir, Deception Reservoir, was built at this time, but it was separated from the SR by a central core rock-fill dam and is no longer used for water supply purposes due to its low water quality. Originally, the dam was located roughly 100 m north of the existing dam and had a spillway elevation of 174.20 masl (first built in 1912). It was upgraded in 1970 and raised to 180.75 masl. In 1992, the intake tower was extended to accommodate the anticipated 2002 expansion. The Council basin is 10.80 km² and lies outside the Sooke watershed boundary, but a diversion from it through Trestle Creek can be used sometime to augment volumes in the SR.
Figure 4-1 - Sooke Reservoir (SR) and surrounding management boundary enclosing most of the Sooke Catchment. Council Creek Basin from which water can be diverted to the SR is outlined. The adjacent Leech Catchment is shown in the inset, along with the location of SR relative to British Columbia.
3 METHODOLOGY AND DATA

The following section describes the methods necessary to derive values for all components of the water balance, and the available data. This section is broken down into total inputs (surface water, diversions, and groundwater) and total outputs (evaporation, spill, and consumption). After each component is quantified, the test of closure for the water balance is outlined and the approach to investigating seasonal trends and extremes is presented.

3.1 Total Inputs

Total inputs to the SR include precipitation on its surface, surface water inflows, and groundwater. The surface precipitation was approximated by the precipitation gauge at Sooke Dam (Fairbairn, 2003). Estimating total surface water inflows to the SR incorporated two techniques: (1) the contributing-area approach; and (2) a hydrologic model. Two estimates were applied, first to allow the methods to be tested against each other to see which was more representative of total inflows to the SR, and second, to understand the strengths and limitations of the approaches. Both required inflow data from streams in the Sooke Reservoir Catchment (SRC) and a Digital Elevation Model (DEM) to which watershed areas could be delineated. Streamflow data were used to calibrate the model to one watershed and to validate its output for other watersheds in the catchment area. Additionally, air temperature and precipitation values were needed to drive the model. Hence, investigations of total inputs were constrained to the PIM. Furthermore, an intermittent diversion into the SR from a catchment area just outside the SR and from groundwater will be discussed later in this section.
3.1.1 Surface Water Inflows

3.1.1.1 Primary Data

Hourly values of precipitation and temperature have been recorded at the toe of Sooke Dam since 1995. Permanently gauged streams, i.e., Rithet and Judge Creeks, drain the two largest catchments in the SR. Both have been gauged since 1993; Rithet with a flume and Judge with a V-notch weir. Their combined gauged area is 42% of the total upland contributing area (26.3 of 63.0 km²). Combined, they comprise 50% of the contributing area to the SR (31.2 of 63.0 km²). Rithet and Judge Creek make up 31% and 19% of the contributing area, respectively, based on the area above the intersection of each creek with the perimeter of the SR. Water-level measurements were made hourly and converted to discharge using rating curves that are regularly calibrated for both high- and low-flow events. Rithet and Judge Creek are also the two largest contributors to the SR by volume. For the period of record (January 1, 1993 to September 15, 2005), Rithet had a maximum discharge of $2.21 \times 10^6$ m$^3$ d$^{-1}$ recorded on September 9, 1999, and a minimum of zero recorded on six consecutive days during September 2002. Maximum discharge in Judge Creek was recorded as $7.11 \times 10^5$ m$^3$ d$^{-1}$ on January 29, 1999. Judge Creek regularly goes dry each year from late June until early November. Although they are both situated in the northernmost slopes of the catchment area, Rithet Creek catchment is west of Judge Creek catchment, has a steeper elevation gain and a larger percentage of high elevation area. Hence, the Rithet Creek catchment tends to receive significantly more precipitation, maintains its soil moisture and tends to support perennial flows. Finally, the 25 km resolution DEM used in the analyses was derived from 1:20,000 maps (TRIM, 2006).
3.1.1.2 Secondary Data

Starting in August 2004, discrete-flow measurements were conducted throughout the watershed to supplement the CRD monitoring program and to understand the ephemeral vs. perennial nature of the various catchments. First approximation stage-discharge curves were developed for the following contributing areas: 17S, Horton, Coquihala, Magee, Whiskey, Maple, Jones, and 3.5 km Creeks (Figure 1). To establish the spatio-temporal pattern of inflows to the SR, three continuous water-level PT2X recording devices were installed in November 2004 on Whiskey, Horton, and Maple Creek. Combining these water-level recording devices provided flow records for an additional 12% of the basin area from November 2004 to November 2005. As a result, 53% of the area of the basin was continuously monitored from November 2004 to November 2005 (Table 4-1).

<table>
<thead>
<tr>
<th>Catchments</th>
<th>Gauged Area (km²)</th>
<th>CRD Composite Analysis Name</th>
<th>Area (km²)</th>
<th>Represents Area (km²)</th>
<th>% Representative Area of Total Catchment Area</th>
<th>% Gauged Area of Representative Area</th>
<th>% Gauged of Total Catchment Area</th>
</tr>
</thead>
<tbody>
<tr>
<td>Rithet</td>
<td>17.75</td>
<td>MAGEE 5.71</td>
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<td>25.03</td>
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<td>71%</td>
<td>28%</td>
</tr>
<tr>
<td></td>
<td></td>
<td>RITHET NORTH 11.33</td>
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<td></td>
<td></td>
<td>RITHET SOUTH 7.99</td>
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<td>Judge</td>
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<td>40%</td>
<td>14%</td>
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<td>3%</td>
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<tr>
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<td>6.63</td>
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<td>63%</td>
<td>7%</td>
</tr>
<tr>
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<td>3%</td>
<td>56%</td>
<td>2%</td>
</tr>
<tr>
<td>Total</td>
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<td>SOOKE RESERVOIR CATCHMENT</td>
<td>62.97</td>
<td>62.97</td>
<td>100%</td>
<td>53%</td>
<td>53%</td>
</tr>
</tbody>
</table>

*Area based on HBV-EC model delineation.

Table 4-1 - Gauged areas and the respective representative areas.

These three streams were selected because they represented the terrain, slope, and aspect types in the catchment area not captured by Judge or Rithet Creeks. Whiskey Creek is
the third-largest contributor to the SR by discharge volume and is representative of streams on the eastern side of the catchment area. It drains a wetland-like area on hummocky terrain, with different features equally distributed among different aspects, and a gentle slope (180 to 630 m). The forest cover in this catchment area has a roughly equal balance of 21-40, 41-120, and 121-250-year-old forests. Horton Creek is representative of streams that drain the west side of the catchment area. It has some of the steepest terrain in the catchment area, gaining over 600 m from the water-level recording device at 170 to 831 metres above sea level at the headwaters, where there is a small lake. Some of the catchment area is non-forested, while other areas are covered with forests 41-120 or 121-250 years old. Most of the catchment area faces east. Maple Creek was selected because spot measurements showed that it had significant discharge during dry summer conditions when most of the other streams were dry. This was likely the result of leakage from the above-located Butchart Reservoir, in spite of a dam separating the two (Figure 4-1). Maple Creek watershed is covered with forest stands 121-250 years old, ranging from 180 to 600 masl. A small portion of the catchment area is covered with stands 0-20 years old. Most of its catchment area faces southeast, then east, and then south, in order of greatest to least area.

3.1.1.3 Methods for Discharge Estimate

Discharge was predicted from water levels recorded on Whiskey, Horton, and Maple Creeks based on the following equation:

\[ Q = \varphi (h - h_o)^\beta \]  \( (m^3 s^{-1}) \)  \( (2) \)

where \( Q \) = discharge \( (m^3 s^{-1}) \), \( h \) is stage \( (m) \), \( h_o \) is a zero stage reference datum \( (m) \) and, \( \varphi \) and \( \beta \) are dimensionless coefficients. The SOLVER tool in EXCEL was used to
minimize the statistic $\sum (\ln Q - \ln Q')^2$ by finding optimal values for the coefficients $\alpha$ and $\beta$ where $Q'$ is the predicted discharge from Eq. (2). Residual plots of $Q$ versus $Q'$ (see Appendix A) were checked for trends to ensure that the equation was not consistently over- or under-predicting high or low flows.

3.1.1.3.1 Modelled Total Inflows

Modelling of total inflows was conducted using the Swedish Hydrologiska Byråns Vattenbalansavdelning (HBV) model, which was first developed in 1972 (Bergstrom, 1995). This model was selected because:

1. It had been successfully applied to watersheds with similar terrain and climate characteristics to those studied in other watersheds in south-western British Columbia; and

2. It represents changes in land cover easily, which makes it useful for conducting long-term studies of the impact of climate and land cover change on inflows (personal communication, Hutchinson, 2004).

The HBV model has the added advantage of moderate data-input requirements, limited free parameters, relative ease of application, and a record of successful application in many parts of the world (Lindstrom et al., 1997; WMO, 1975 and 1986). It is a semi-distributed, deterministic model. Sub-basins form the primary hydrological unit within which similar area-elevation bands are defined and a crude classification of land use (forest, open, lakes) is made (Bergstrom, 1995). Modifications to the code in 2000 created the HBV-EC, which is Environment Canada’s variant of the model (Moore, 1992; Hamilton et al., 2000). This version allows multiple climate zones to be defined within one catchment area and further “discretization” based on elevation, slope, and aspect. A variable reservoir configuration (parallel vs. serial) was added to this version and applied in this study.
Model calibration was performed to ensure that the estimated representative factors for rainfall and snowfall and their respective correction factors (RFCF and SFCF) produced reasonable volumes of catchment area runoff. Next, the parameters guiding the runoff routines were adjusted to fit runoff to timing of the observed hydrograph. These parameters are the runoff fraction to fast reservoir (FRAC), the fast reservoir coefficient (KF), the fast reservoir exponent (Alpha), and the slow reservoir coefficient (KS). The other free parameters were then adjusted to obtain suitable magnitudes and timing. Calibration was conducted by trial and error, visualizing the modelled versus observed discharge values for the basin. The results were evaluated using the Nash and Sutcliffe (1970) $R^2$ (NASH) statistical criterion. Ideal NASH results have $R^2$ values close to 1.0. Those above 0.8 are considered high, and those below 0.6 are considered to be less than suitable and suggest that the model should be adjusted or the observational data should be further verified. The Mean Volume Error (MVE) was used to indicate the strength of the calibration with ideal values being between zero and 10%. The final calibration maximized the NASH and minimized the MVE, thereby taking into consideration both objective functions. This approach has been successful in calibrating the HBV-EC model in other watersheds in southwest BC with similar characteristics (personal communication, Hutchinson, 2006). The model was run from January 1, 1993 to September 19, 2005 at a daily time step for Rithet and was calibrated visually to match the timing and magnitude of the observed discharge from October 1, 1993 to September 19, 2005. This provided a 10-month lead-up to the start of the 1993-1994 water year. The observational data available included a variety of hydrological events and was of adequate length to allow for all the subroutines of the model to be calibrated.

To provide validation, the same values of the parameters that were defined by calibration on Rithet Creek were applied to Judge Creek, altering only the RFCF and SFCF factors (from 1.1 to 0.96; see Figure 4-2) as Judge Creek is known to receive less precipitation than Rithet Creek (Fairbairn, 2003; Niemann, 1993). The model was run on a
daily time step and simulated discharge for Judge Creek, which was visually compared to the
daily observed discharge to test its ability to replicate the timing and magnitude of the
observed discharge. The fit was evaluated with the *NASH* and *MVE* scores. To further
validate the model, the same calibrated values for the parameters were applied to 17S,
Horton, Magee, Coquihalla, Whiskey, Maple, Jones, and 3.5 km Creeks. The *RFCF* and
*SFCF* were the only parameters that were given different values, which were 1.1 for Magee,
Horton, and 17S Creeks, and 1.0 for Whiskey, Maple, Jones, and 3.5 km Creeks. Magee,
Horton, and 17S Creek are on the western boundary of the catchment area, which is closer to
the ocean and at a higher elevation, and therefore receives more precipitation than the Sooke
Dam station. Whiskey, Maple, Jones, and 3.5 km Creek were designated 1.0 to reflect their
position on the north-eastern boundary of the catchment area, which receives roughly the
same amount of precipitation as Sooke Dam (Fairbairn, 2003; Niemann, 1993). The model
was then run on an hourly basis for all the basins. Where instantaneous discharge
measurements were available for the (2004-2005) period, outputs were visually compared
against these values for timing and magnitude to see how closely the spot values fell on the
line predicted for streamflow by the model.

3.1.1.3.2 Scaling Inflows from Individual Catchments up to the Total Inflow Estimate

The inflow volume from each modelled catchment area was then converted to its
representative volume by summing the area of catchments that had similar physical
characteristics to get areas for which modelled inflow would be representative. The inflow
volume from the modelled catchment area was then multiplied by the ratio of the
representative area versus the modelled area to get the representative volume. The
relationship is defined in the following equation:
\[ Q_2 = Q_1 \left( \frac{A_2}{A_1} \right) \text{ (m}^3 \text{ s}^{-1}) \]  

where \( Q_1 \) = discharge from the modelled area (m\(^3\) s\(^{-1}\)), \( A_1 \) = area that is modelled (m\(^2\)), \( A_2 \) = area of the representative area (m\(^2\)), and \( Q_2 \) = total discharge from the representative area (m\(^3\) s\(^{-1}\)). The selection of the similar areas was based on the slope, aspect, and land cover types classified by the CRD in their 2001 Compartment Analysis of the entire water-supply area (CRD 2001). Modelled catchment areas were delineated using the HBV-EC model. The catchment area above the intersection between the stream and main road were selected for the majority of the catchments because discharge measurements were made at the cross-section in the creeks nearest the road for these locations. The areas for Rithet and Judge Creeks were defined by the catchment area above the weirs respectively (see Table 4-1).

The total inflow from each representative area was then summed to give the combined total inflow from the entire catchment, which equalled 100% of the total 63.0 km\(^2\) catchment area for the HBV-EC modelling approach (Table 4-2). The total inflows estimated by this approach are hereby denoted as \( I_\eta \) (m\(^3\)).

3.1.1.3.3 Estimates of Total Inflows Based on Contributing Area Approach

Total inflow from the catchment area to the SR was estimated using the contributing-area approach. Like the method described for the modelling approach, for the contributing-area approach, inflow volumes from each catchment were converted to representative volumes using Eq. (3). With the contributing approach \( Q_1 \) became the discharge from the gauged area (m\(^3\) s\(^{-1}\)) (instead of modelled) and \( A_1 \) became the area that was gauged (m\(^2\)) (instead of modelled). \( Q_1 \) was then multiplied by the ratio of representative area to gauged area where the representative areas were re-selected to make Rithet Creek and Judge Creek inflows representative of a greater proportion of the total catchment area than they were.
previously representative of in the modelled approach (see Table 4-2). The representative areas were again selected based on the physical catchment descriptions provided in the CRD 2001 Compartiment Analysis (CRD, 2001).

<table>
<thead>
<tr>
<th>Catchments</th>
<th>Modelled Area (km²)</th>
<th>CRD Composite Analysis Name Representative</th>
<th>Area (km²)</th>
<th>Representative Area (km²)</th>
<th>% Representative Area of Total Catchment Area</th>
<th>% Modelled Area of Representative Area</th>
<th>% Modelled Area of Total Catchment Area</th>
</tr>
</thead>
<tbody>
<tr>
<td>Rithet</td>
<td>17.75</td>
<td>RITHET NORTH</td>
<td>11.33</td>
<td>19.32</td>
<td>31%</td>
<td>92%</td>
<td>28%</td>
</tr>
<tr>
<td></td>
<td></td>
<td>RITHET SOUTH</td>
<td>7.99</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Judge</td>
<td>8.54</td>
<td>JUDGE</td>
<td>11.87</td>
<td>11.87</td>
<td>19%</td>
<td>72%</td>
<td>14%</td>
</tr>
<tr>
<td>Horton</td>
<td>1.64</td>
<td>HORTON</td>
<td>7.74</td>
<td>7.74</td>
<td>12%</td>
<td>48%</td>
<td>6%</td>
</tr>
<tr>
<td>17S</td>
<td>2.07</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Magee</td>
<td>0.93</td>
<td>MAGEE</td>
<td>5.71</td>
<td>5.71</td>
<td>9%</td>
<td>45%</td>
<td>4%</td>
</tr>
<tr>
<td>Coquihalla</td>
<td>1.64</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Jones</td>
<td>2.19</td>
<td>TRESTLE &amp; ACCESS</td>
<td>4.95 &amp;</td>
<td>9.68</td>
<td>15%</td>
<td>30%</td>
<td>5%</td>
</tr>
<tr>
<td>3.5 km</td>
<td>0.69</td>
<td></td>
<td>4.73</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Whiskey</td>
<td>4.16</td>
<td>WHISKEY</td>
<td>6.63</td>
<td>6.63</td>
<td>11%</td>
<td>63%</td>
<td>7%</td>
</tr>
<tr>
<td>Maple</td>
<td>1.13</td>
<td>Maple*</td>
<td>2.02</td>
<td>2.02</td>
<td>3%</td>
<td>56%</td>
<td>2%</td>
</tr>
<tr>
<td>Total</td>
<td>40.74</td>
<td>SOOKE RESERVOIR</td>
<td>62.97</td>
<td>62.97</td>
<td>100%</td>
<td>65%</td>
<td>65%</td>
</tr>
</tbody>
</table>

*Area based on HBV-EC model delineation.

**Table 4-2 - Areas as modelled by HBV-EC and the respective representative areas.**

An important component this approach is that flows from Maple, Whiskey and Horton Creeks were estimated based on regression relationships established between Rithet Creek and Maple Creek, Rithet Creek and Horton Creek, and Judge Creek and Whiskey Creek (Appendix B) based on data collected during the November 2004 to November 2005 monitoring period (see Section 3.1.1.3). Discharge from Maple, Whiskey, and Horton Creeks was measured for only this thirteen-month window out of the total 108 months in the PIM. Hence, this method was used to incorporate the information gained during the thirteen-month monitoring period to estimate discharge from these catchments for the PIM.
The inflows from each representative area were again summed to arrive at the total inflow from the total 63.0 km² or 100% of the catchment area for the upland contributing-area approach. The total inflow volumes estimated with this approach are denoted as $I_X$ (m³).

3.1.2 Diversions

Council Creek Reservoir is often used to augment SR volumes and therefore complete calculation of inflows requires accounting for water attained via intermittent diversion from this reservoir. In 1992, a diversion was created to draw from Council Creek (Figure 1) into the SR through Trestle Creek, at the most south-eastern point of the SR (personal communication, Gudavicius, 2006). Discharge measurements have been made since 1992 with a V-notch weir on the outflow of Council Creek. Prior to October 1, 1995, these readings were taken every two hours and subsequent readings were taken every hour. Volumes drawn to the SR are estimated as a percentage of the volumes measured at the outflow of Council Creek; 10% when the valve is fully open and 5% when the valve is partially open (personal communication, Gudavicius, 2006). There are limitations set on the daily and annual total flows that can be diverted from Council Creek Reservoir into the SR.

3.1.3 Groundwater

Kenney (2005) investigated aquifer types in the CRD and documented areas near to the SR. In this study, it was found that most of the materials in the area had discharge on the order of $10^2$ - $10^3$ m³ month⁻¹. From this, one can expect the groundwater contribution to the SR to be minimal relative to the other terms. Furthermore, the SR is situated below 800 m hills to the west and 500 m hills to the east. Standing bodies of water are located at the top of both higher elevation areas and could be a source of recharge to groundwater that feeds into the SR. However, the SR is located up-gradient of the Sooke River valley, which creates
potential for it to be a discharge zone to those elevations below it. Therefore, input and output volumes of groundwater may balance out.

A first-order approximation of possible groundwater discharge into the SR is estimated based on monthly discharge measured at Rithet when Judge Creek had zero discharge. The discharge from Rithet Creek during these times is likely to be representative of baseflow provided by groundwater sources. These low-water periods usually occur in August and September and sometimes as early as July. The total discharge from Rithet Creek is divided by the drainage area of the Rithet basin (17.75 km²) to get depth of discharge. Prior to this, any precipitation falling on the catchment area will be subtracted from the volume of flow. The resulting depth of discharge over the catchment area is then multiplied by the average surface area of the SR to estimate total groundwater inputs into the SR, and is expressed in units of contribution in m³ month⁻¹.

A more detailed monitoring or modelling of groundwater is beyond the practical scope of this study, and it is therefore assumed that the estimated values of $I_G$ and $O_G$ are reasonably small compared to the magnitude of the other water-balance terms. This first-approximation analysis will provide some evidence to support or refute this assumption.

3.2 Total Outputs

Outputs from the SR comprise evaporation, spill, fisheries release, and consumption. The following discusses the methods used to measure or estimate these components of the water balance.

3.2.1 Evaporation

Estimation of evaporation required detailed analysis and is reported in a companion report (Werner, 2007b). As background, the Penman (Penman, 1948) approach was used to estimate evaporation. This method has been found to produce values that match those of the
energy-budget method more closely than any other (Winter et al., 1995). Net radiation, air temperature, humidity, wind speed, and heat storage in the lake have been measured at regular intervals since October 1996. These data allowed this relatively detailed approach for estimating evaporation to be applied and evaluated for response to climate conditions over a broad range of values during the PIM. Further details can be found in Werner (2007b).

3.2.2 Spill

Water is spilled from the SR during high-water events when water levels exceed the capacity of the dam. This typically takes place in the winter months of January, February, and March. During the dry season, water-levels are low and it is not necessary to spill excess water, so during these times, spill does not need to be considered in the water-balance accounting. Modifications to the dam for the purpose of expansion, including those in 1970 and 2002, have resulted in modifications to the spillway location and type. A new concrete gravity-free crest spillway was built and connected to an earthfill embankment for the 2002 upgrade. This 100 m-long embankment connects the new spillway to the left abutment. Spill measurements up to and including 1998 were calculated based on only one daily water-level reading. Starting on December 6, 1998, hourly readings of water-levels were made and used to calculate an average daily spill volume (personal communication, Gudavicius, 2006).

3.2.3 Fisheries Release

To provide reliable downstream flow for in-stream fisheries, controlled releases began in February 2004 in response to an agreement between the CRD, the Department of Fisheries and Oceans, the B.C. Ministry of Water, Land and Air Protection, and T'Sou-ke First Nation. Control gates were incorporated into the spillway to allow release of water to the Sooke River in accordance with the agreement. Fisheries releases after February 2, 2004 are based on actual water released and measured with a magnetic flow meter using the same
method described below for consumption. Prior to February 2, 2004, any fisheries releases from SR were estimated because there were no flow records (personal communication, Gudavicius, 2006). A new concrete-lined spillway was constructed on the right abutment of Deception Dam, discharging into Deception Creek just below the dam in 2002 (CRD website, 2006). Water from the Deception Reservoir is sometimes used to supply the Sooke River.

3.2.4 Consumption

Consumption amounts are driven by demand. Consumption in the summer is double that in the winter because of increase demand of water for gardening and other outdoor activities (based on CRD data from 1996 to 2005). Higher temperatures tend to increase this effect. To control water use in summer, three stages of water restrictions have been developed: stage 1 is moderate, stage 2 is acute, and stage 3 is severe (CRD, 2006). Stage 1 is automatically implemented every year from June 1st to September 30th, which limits the household watering to two days per week and aims to reduce overall water use by 10%. In 2001, rainfall was the lowest since 1900 and the SR was 30% lower than normal for that time (Walker, 2002). Stage 3 water restrictions were put in place to reduce water use by 25-30% of normal amounts for the summer period. During this time, no outdoor water use was permitted; lawns could not be watered, fountains could not run, pools could not be refilled, and cars could not be washed (CRD, 2004a). Stage 2 is an intermediate restriction that has rules that fall between those from Stages 1 and 3.

Water is drawn from the lake for outflow via the intake tower through one of three sets of 1.5 m x 1.5 m gates. The water is then conducted through the outflow pipe where the volume of outflow is measured. Measuring practices have changed over the period of record. Prior to 1970, when the new Sooke Dam was built, outflow was measured using a weir to determine water released at Sooke Dam. Although official records are not available, the water released at Sooke Dam was likely set on a daily basis and did not change throughout the day.
Between roughly 1970 and 1992, outflow was measured by a mechanical meter that was read once per day (specific dates are not known). After 1992, venturi meters connected to a mechanical totalizer were used and read once daily. After 2002, outflow was measured by exception using a magnetic flow meter with a SCADA data logger that records flow by exception roughly every minute (personal communication, Gudavicius, 2006). Errors in measurements of flow are estimated to be ±5%.

### 3.3 Change in Storage

The volumes of change in storage in the SR are estimated based on a relationship of water levels to volume provided by the CRD. This relationship was used to define a polynomial equation relating volume to water level. The range over which the relationship applies is 113.6 masl to 190.0 masl. Accurate assessment of the volume is dependent on the bathymetry of the SR basin as evaluated by the CRD. Water-level measurements up to and including 1998 were made once per day. Starting on December 6, 1998, hourly readings of water levels were taken with an automated water-level recording device (personal communication, Gudavicius, 2006).

### 3.4 Closure of the Water balance

The overall accuracy of the water-balance calculations is assessed by comparing the calculated value of $\Delta S$ from Eq. (1) to the observed change in storage obtained from a regression between volume and water levels. The change in storage based on difference in volumes is defined as $\Delta S_{\text{obs}}$. Closure will be investigated via cumulative distribution plots.

### 3.5 Seasonal Trends and Extremes

To assess multi-year trends in water-balance conditions, the Palmer Drought Index (PDI) (Palmer, 1965) was applied to temperature and precipitation collected at Sooke Dam
The PDI has been used globally (Dai et al., 2004), including the United States and Canada (Akinremi et al., 1996), to assess the severity of past and ongoing droughts and wet spells. First developed in 1965, the PDI is a measure of long-term cumulative meteorological drought and wet conditions based on records of precipitation and air temperature and an approximation of soil type. Temperature and precipitation records collected at the Sooke Dam from 1916 to 2005 were used to compute the PDI for the SR. The results of the PDI were used to place the worst-case drought scenarios in the PIM in the context of the longer-term record. Additionally, the PDI provided perspective on the relative magnitude of dry and wet events within the PIM compared with those in the long-term record.

Monthly values for the water-balance components were compiled into six-month seasons, wet (October to March) and dry (April to September) to enable seasonal comparisons. To derive potential worst-case scenarios of drought for the SR supply, the wet season with the lowest precipitation was combined with the dry season that had the highest evaporation. All components of the water balance for the selected seasons were used in computing the annual water balance for the worst-case scenario. For example, the actual evaporation and inflow magnitudes that had occurred during the wet season with the lowest precipitation were included. Adjustments were made, however, to the amount of consumption to reflect likely values of consumption during the worst-case scenario. Consumption values from the 2001 dry season were used because this was the only time on record when Stage 3 water restrictions were put in place (see Section 3.2.4). Such restrictions would be put in place during extreme drought conditions. Computations were made in units of water level (masl) to eliminate the inclusion of intra-annual surface area variation, which would prevent direct comparison of changes in storage from season to season.

Two scenarios were investigated, each with different initial conditions. In the first scenario, water levels in the SR were set at the average water level for April 1st, and in the
second they were set at the maximum water level for April 1<sup>st</sup> over the 2003 to 2005 period. Starting from the initial water level (masl), water was added for the wet season and then subtracted for the dry season until the levels in the SR were below 170 masl. This 170 masl-level is defined as the threshold below which water in the SR is considered to be inaccessible or “dead storage.” This provided an estimate of how many years it would take before the SR would run into storage volumes that could not be accessed via free drainage into the intake tower.

4 RESULTS AND DISCUSSION

In the following section values for results for all components of the water balance are described and discussed, including total inputs (surface water, diversions, and groundwater) and total outputs (evaporation, spill, and consumption). Closure of the water balance is discussed and seasonal trends and extremes are presented.

4.1 Total Inputs

4.1.1 Surface Water Inflows

4.1.1.1 Validity of HBV-EC Model

The $NASH - MVE$ values for observed to calculated discharge for Rithet and Judge Creeks were 0.79 – 9.84% and 0.78 – 9.72% respectively over the 1995 to 2006 period. These values indicate that the model simulates the discharge for these two catchments well, and that a model calibrated to one catchment area can be applied to a nearby catchment area for the same period of record with reasonable results. Based on visual inspection of the plots of daily observed modelled flow for Rithet and Judge Creeks (Figure 4-2 and Figure 4-3), it can be seen that accurately modelling the timing of the transition from the dry to the wet period is more problematic for the ephemeral Judge Creek than for the perennial Rithet Creek. This
suggests that the model is not capturing some of the physical characteristics of Judge Creek that cause it to act ephemerally. Close mirroring of observed and modelled hydrographs for the other catchments (17S, Horton, Coquihalla, Magee, Whiskey, Maple, Jones, and 3.5 km Creeks) testifies to the validity of the HBV-EC results (not shown).

Figure 4-2 - Observed discharge measured at the Rithet Creek weir compared to the discharge simulated with the HBV-EC model ($x10^5$ m$^3$ day$^{-1}$) over August 1997 to September 1999.
A residuals plot of simulated minus observed discharge, versus observed discharge for Rithet Creek (Figure 4-4) shows that most of the differences are spread equally above and below zero. When observed values are greater than $8.59 \times 10^5 \text{ m}^3 \text{ d}^{-1}$, the simulated values consistently underestimate the observed values. This indicates that the model is missing the peaks of the large discharge events. The $NASH$ function is normally known to display a larger sensitivity to flood than to base flow conditions because it is a function of the sum of squared errors (Liden, 2000). In the SRC, peak flows are event based, with flows quickly returning to low-flow conditions after each event. In this type of system, the strength of the $NASH$ value will be weighted more strongly towards measuring the ability of the model to simulate these low-flow events because of their prevalence over the year.
4.1.1.2 Comparison of Total Inflow Estimates

As shown in Figure 5 in almost all cases, the HBV-EC model estimates of total inflow ($I_\eta$) were less than those from the contributing area approach ($I_X$), as shown in Figure 4-5. This is likely due to the aforementioned underestimation of peak flow and differences in low flow methods. In particular, two factors contributed to $I_\eta$ being less than $I_X$. The first is that the peaks were compromised (as described above) and this effect was then multiplied across the basin. The second is that the low flows are less in the HBV-EC model estimate than they are in the contributing area approach because of differences in the way the two methods assess flows from Maple Creek. Low flows over the dry season (April to September) may be overestimated in the contributing area approach because they are primarily constituted by streamflow from Maple Creek, because the other creeks are generally dry.
Figure 4-5 - Comparison of inflow estimates from the HBV-EC model and contributing area approaches (m³ day⁻¹).

Flows from Maple Creek are anomalously large because during the dry season due to a leak from the Butchart Reservoir. Discharge averaged $1.80 \times 10^5$ m³ d⁻¹ over the 13 month period it was monitored. Streamflow was measured for year only and therefore it cannot be expected to have the same relationship to Rithet Creek flows each year. Amounts would be dependent on the pressure head created by the water level in the Butchart Reservoir, but there is not enough data on water level in the Butchart Reservoir, or on discharge from Maple Creek, to assess the relationship between the two. Furthermore, it would be difficult to adjust the HBV-EC model to represent this leak from the Butchart Reservoir into Maple Creek.

This comparison has demonstrated that the two methods of estimating inflows mirror each other in timing and have slight differences in magnitude. This study incorporates inflows estimated using the contributing-area approach since this method was more successful in estimating the peak-flow conditions than the modelling approach. These peak
flows were found to contribute a large percentage of the monthly flow volumes. Therefore, the contributing area approach yields higher magnitude inflows that likely approximate the true inflows more accurately.

4.1.1.2.1 Timing and Magnitude of Flows

Inflows have strong inter- and intra-annual variability. Peak monthly values occur most often in December or January, and minimums span from July through October (Figure 6). The largest volumes in the record occurred during the 1998 to 1999 wet period. This period was affected by La Nina conditions (Kiffney et al., 2002; Flemming et al., 2006) that were coupled with a cool phase of the PDO. This combination has been shown to bring wetter than normal conditions to the southwest of BC (Kiffney et al., 2002). Alternatively, the preceding dry season, April to October 1998, was one of the strongest El Nino events on record, and was coupled with a warm-phase PDO (Kiffney et al., 2002). Water volumes in the SR during the dry season of this year were the second lowest on record after those in 2001 (Figure 4-6). Inflows in April, May and June 1998 were also low relative to other years, adding to the deficit.
During the (2000-2001) water year, inflows were the lowest recorded. Precipitation during this year was the lowest measured for the last 100 years (Walker, 2002). However, this was not an El Niño year, which is usually associated with low precipitation and high evaporation in this region. Evaporation occurred at normal levels, which suggests that evapotranspiration from the SRC was also similar to other years except for being limited by a potentially drier landscape. With low precipitation and relatively strong evaporation, the resulting inflows were reduced. The cause for low precipitation in this year has not been established.

A quick evaluation of the relationship of runoff (R) to precipitation (P) shows that R/P is not uniform from year to year, suggesting that variation in evapo-transpiration rates in the catchment area significantly alter runoff. Hence, evaporation estimates from the lake cannot be used to estimate evapo-transpiration directly. In addition, the spatially varying
vegetation coverage and radiation regimes due to different aspects prevent clear estimates of evapo-transpiration across the SRC.

4.1.2 Diversions

As described in the methodology, Council Creek is a resource that can be drawn on to augment SR volumes. Initial estimates of discharge at Trestle, the stream through which Council flows are diverted to the SR, are different from what would be expected from the measurements at the weir on the outflow of Council. The volumes at Trestle are sometimes an order of magnitude larger than those estimated at the weir. Discrepancies could arise from a number of factors including flows from the Trestle catchment area contributing to the total volume and rating curve relationships for the weir on the Council Creek diversion not being tested for high and low events. Most importantly, this weir is said to not be set up in a stable cross-section (personal communication, Miles, 2006). Re-evaluating the weir set-up and reassessing the rating curve would contribute to improved confidence in the estimate of flow from this diversion to the SR. Based on the data provided the magnitude of this diversion is often less than 10% of total inflows from the SRC, however, give the state of the weir error in this term could be substantial, which would contribute error to estimating total inflows.

4.1.3 Groundwater

Baseflow conditions in Rithet can range from $3.15 \times 10^4$ m$^3$ month$^{-1}$ to $1.85 \times 10^5$ m$^3$ month$^{-1}$ when Judge Creek has no discharge. The median value for discharge is $6.95 \times 10^4$ m$^3$ month$^{-1}$ and the mean is $7.98 \times 10^4$ m$^3$ month$^{-1}$. Converting this mean value into the depth of runoff from the catchment area yields $4.50 \times 10^3$ m month$^{-1}$. Computing the amount entering the SR from all sides of the SR yields $2.67 \times 10^4$ m$^3$ month$^{-1}$. Groundwater inflows values are two orders of magnitude smaller than inflows estimated from the whole catchment area on average annually, and are less than 1% of the total volume of water estimated to be lost to
evaporation over the summer months. Based on this analysis, the groundwater flow rates are (a) likely to be insignificant relative to the other terms in the water balance; (b) likely fit within the error for the overall budget; and (c) can be safely assumed to balance out to zero net gain to the SR. The possibility and affect of more localized or regional flow along fracture zones, known to be present in the area of the SR, has yet to be evaluated.

4.2 Total Outflows

4.2.1 Evaporation

The average monthly evaporation over the nine-year study period was 67 mm. There was a large standard deviation of 51 mm and coefficient of variation of 1, which demonstrates the strong seasonality of the evaporation rates. The maximum monthly total evaporation of 194 mm occurred in August 1998, and the minimum of 0.2 mm took place in January 2004. Table 4-3 provides a summary of the average monthly values. Monthly evaporation is lowest in March at 12 mm on average, and highest in August at 158 mm on average. The standard deviation is highest for the months of June and August, at 27 and 24 mm per month respectively. The lowest standard deviations were found to occur in February and March, at 6 mm per month, respectively. Annually, the highest evaporation years were the 1997-1998, 2001-2002, and 2002-2003 water years.
<table>
<thead>
<tr>
<th>Month</th>
<th>Mean (mm)</th>
<th>Min. (mm)</th>
<th>Max. (mm)</th>
<th>Std. Dev. (mm)</th>
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</thead>
<tbody>
<tr>
<td>October</td>
<td>65</td>
<td>37</td>
<td>80</td>
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<tr>
<td>November</td>
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<td>28</td>
<td>10</td>
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<td>28</td>
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<td>May</td>
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<tr>
<td>August</td>
<td>158</td>
<td>115</td>
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<td>24</td>
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<td>September</td>
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</tr>
<tr>
<td>Average</td>
<td>63</td>
<td>42</td>
<td>83</td>
<td>13</td>
</tr>
</tbody>
</table>

Table 4-3 - Monthly evaporation values of mean, minimum, maximum, and standard deviation (mm).

4.2.2 Spill

Prior to the raising of the dam in 2002, it was necessary to spill water during high water events when the capacity of the dam had been exceeded. These spill events took place only during the months of December to April. During the dam construction, and for a few months following the completion of the dam, water was spilled purposely to prevent the SR from filling too quickly. Spill was necessary in small amounts in January, February, March, and October 2004 when the SR had reached full capacity and could not take on any more water.

4.3 Change in Storage

Observed change in storage of the SR has a distinct, re-occurring seasonal pattern. Precipitation generally starts in late October or early November and reservoir storage has a primary peak in December or January, followed by an intermediate low that is followed by a
secondary peak in March or April. After this point, the SR starts to lose more water through continuous consumption than it gains through precipitation and inflows.

Four of the nine water years shown have patterns that are different from the average over the PIM. The (1997-1998) and (2000-2001) water years had significantly less storage during the primary influx period. For the (1997-1998) water year, less water arrived during the October-January period, the common influx period, but for the (2000-2001) water year the start of the influx, or the increase in storage, arrived much later than other years. It did not start until December, and the secondary input in March-April was limited. The (2002-2003) water year had the largest second peak over the February-March period of any of the other years. This could be attributable to higher-than-average inflows in March 2003, and to the completion of the raising of the dam in December 2002, which increased the capacity of the SR, and allowed for more storage than previous years where excess would have been spilled.

The 2003-2004 water year showed only one peak, rather than two. This year had a unique input of water via what is commonly referred to as a “pineapple express,” which is a Pacific Ocean subtropical jet stream that brings warm moist air from Hawaii to the west coast of North America (Stahl et al., 2006). This event raised water-levels in the SR by 1 m in less than 24 hours in October 2003. This extreme high-water event was followed by less-than-average inputs in March-April 2003 in comparison to other years in the PIM. Although “pineapple express” events took place in other years during the PIM, this one was the largest.

The minimum water level for the PIM was 173.7 masl on November 1, 1998; the maximum was 186.5 masl on May 1, 2005; the average was 179.9 masl; and the standard deviation was 3.3 masl. Although the lowest water-level was reached during the 1998-1999 water year, the water year with the lowest annual water-level was 2000-2001, suggesting that the dry period was longer in this year than it was in the 1998-1999 year. The maximum annual average water level was reached in the 2004-2005 water year. On average, monthly
water levels peaked in April (182.3 masl) and reached their minimum in October (176.4 masl). There is a standard deviation of 2.2 masl for the monthly averages.

Simulated volumes match observed values well, by ±1% on average. Underestimates were made in the wet season, but during the dry season, simulated values match the observed closely. Figure 4-6 shows annual and seasonal trends in SR volume. Pre- and post-December-2002 thresholds of maximum capacity are evident in this figure.

4.4 Test of Closure of the Water balance

Change in storage was computed by adding the inputs and subtracting the outputs of the water balance for each month. These monthly values were added to each other starting with October 1996 through to September 2005 to devise the cumulative distribution plot of $\Delta S$ versus $\Delta S_O$ (Figure 4-7). Over this 108 month period, $\Delta S_O$ gains and losses water during the year, but for the most part stays above zero. Conversely, water is gained and lost in $\Delta S$ over the year; however, the cumulative values quickly start to decline below zero. By September 2005, the difference between $\Delta S$ versus $\Delta S_O$ is $1.62 \times 10^8$ m$^3$. The rate at which the cumulative $\Delta S$ becomes negative is greatest before October 2000. After this month, the shape of $\Delta S$ starts to match that of $\Delta S_O$, except for 2001-2002, when $\Delta S$ becomes more negative before flattening out again in late 2002.
The observed storage and the change in storage computed by adding the inputs and subtracting the outputs of the water balance for each month was plotted as a time series (Figure 4-8). For most of the record, the computed storage matches the observed storage closely, suggesting that the individual components of the water balance are well accounted for. The 2000-2001, 2002-2003, 2003-2004, and 2004-2005 water years match most closely. The 1996-1997, 1997-1998, 1999-2000, and 2001-2002 water years show discrepancies during the January-March period, but match closely during the other months of the year. Overall, large discrepancies seem to be isolated to the wet season, especially to January. Alternatively, May to September consistently shows good closure of the water balance. These discrepancy events align temporally with spill events, and seem to have a proportional magnitude to them.
Figure 4-8 - Change in storage based on total inputs (precipitation + inflows) minus the total outputs (evaporation + outflow) in the water balance, observed change in storage, and spill (m³).

Over-estimates of the volume of water spilled and of the volume of water stored in the SR could account for the difference between $\Delta S$ and $\Delta S_O$. As both spill and $\Delta S$ are estimated based on water level, they are both affected by imperfect measurements of water level; however, their errors act in opposition to exaggerate the difference between $\Delta S$ and $\Delta S_O$. To illustrate, if the water-level was measured as 186 masl when it was truly 185 masl, the spill would have been over-estimated, which would have resulted in an over-estimate of volume lost and a greater-than-actual negative $\Delta S$, while $\Delta S_O$ would be a greater-than-actual positive value because the water-level increase was overestimated. The under-estimate of $\Delta S$ and $\Delta S_O$ over-estimate leads to large differences between the two values.

Errors in spill values are potentially significant. Figures 4-9 and 4-10 compare daily spill estimated from daily water-level records to those estimated from hourly water-level readings. Spill estimates are dependent on water elevation in (Figure 4-9), but are not in
Both relationships are exponential, increasing the effect of inaccurate water-level estimates on spill estimates. The relationship of SR volume to water level is similarly exponential.

Also of note is that water-level measurements were made only to the second decimal place (i.e., 0.01 m) before December 6, 1998, and after December 6 they were made to the third decimal place (i.e., 0.001 m). Depending on the water level in the SR, differences in the level of precision could account for three orders of magnitude differences in spill volume. For example, the difference would be $6.95 \times 10^3$ m$^3$ if the measurement were recorded as 184.001 m versus 184.00 m. Thus, spill estimates made before December 1998 are likely to have error, which makes it difficult to close the water balance during months when spill took place.

![Graph](image-url)

**Figure 4-9** - Daily spill estimates (m$^3$) based on daily water elevation measurements.
Each term in the water balance has error and contributes to some degree to these discrepancies even if water level is the primary factor. The true nature of the discrepancies is difficult to define when values are summed over a monthly basis. Evaluating the water balance daily revealed that the timing of the large differences coincided with large inflow events. It also coincided with high water levels in the SR that occurred just before the flashboards were installed (this is when the head over the spillway is at its greatest), and discrepancies were demonstrated to be magnified when large inflow events coincided with spill events.

### 4.5 Water Balance Partitioning

The annual water balance of the SR averaged over the PIM showed that total inputs are dominated by $I_S$ (89%) and supplemented by $P$ (11%). Similarly, outputs are heavily dominated by $O_S$ (96%) with only 4% leaving via $E$ (Figure 4-11). There is a large range in
annual values of $I$ and $P$, but the variability in $I$ is much greater than for $P$, suggesting that evapo-transpiration rates vary from year to year. A preliminary analysis of Runoff/Precipitation (R/P) ratios showed significantly annual variability. The variation in estimates of annual evaporation from the SR supports the hypothesis that there is variation in annual evapo-transpiration. Further details of these results are provided by Werner (2007b).

![Figure 4-11 - Partitioning of the annual water balance into total inflows, precipitation, outflows, and evaporation (%).](image)

The relative importance of the water-balance terms changes significantly from the wet season to the dry season. Over the wet period, outputs are primarily $O_S$ (99%) and secondarily $E$ (1%), and inputs are made up of $P$ (10%) and $I_S$ (90%). This contrasts with the dry period when outputs are a sum of $O_S$ (91%) and $E$ (9%), and inputs are contributed by $P$ (22%) and $I_S$ (78%; Figure 4-12). There is increased evaporation/evapo-transpiration in the dry season, which increases the ratio of $E$ versus $O_S$ and reduces the ratio of $I_S$ versus $P$ when compared to the wet season. The relative magnitude of $P$ versus $E$ varies strongly by season. During the wet season the depth of $P$ is 8.9 times the depth of $E$; conversely, during the dry season the depth of $E$ is 2.7 times the depth of $P$. 
Figure 4-12 - Partitioning of the water balance into total inflows, precipitation, outflows, and evaporation (%) for the wet (October – March) and the dry (April – September) seasons.

Comparison of the P-E index for the wet versus dry period further confirms the seasonality of the water balance and quantifies the magnitude of the difference in these values (Figure 4-13). The volume of water surplus during the wet period is often five times greater in magnitude than the dry-period deficit. This high-intensity water input over this six-month window renders those volumes that are beyond the capacity of the SR storage unusable, forcing them to be spilled. The wet period always shows a water surplus and the dry period always shows a water deficit, except for the April-September 1997 dry season, when there is a small magnitude surplus. This year had more than twice as much rain as most of the other dry seasons over the period of study.
Figure 4-13 - Precipitation minus evaporation (mm) for each wet (October-March) and dry (April-September) season over the study period (October 1996-September 2005).

\[ O_S \] is composed of consumption \((O_{S1})\), spill \((O_{S2})\), and fisheries release \((O_{S3})\). \(O_{S2}\) occurs only during the wet season. \(O_{S1}\) is lower in the wet season than in the dry season, when demand for water increases for landscaping and other outdoor uses. \(O_{S1}\) has been shown to be greatly adjustable with Demand Side Management Practices that were described in Section 3.2.3.

### 4.6 Seasonal Trends and Extremes

In addition to the strong contrasts in moisture balance from the wet to the dry season, there is also significant inter-annual variation of the water-balance components over the PIM. Precipitation and inflow amounts fluctuate more than those for evaporation and outflow from year to year over the wet season. Evaporation rates are most influential on the water balance in the dry season. Strengthening of evaporation during the dry season tends to coincide with less rainfall occurring over the dry season, which when combined, maximizes the water
deficit. Over the PIM, the wet season with the lowest precipitation amount was October 2000 to March 2001 with 703 mm of precipitation; the dry season with the highest evaporation was April to September 1998 with 706 mm of evaporation.

The PDI was computed based on monthly data from 1919 to 2005 (Figure 4-14). In these records, the (2000-2001) wet season has large negative values relative to other years. However, there are times in the past when the PDI has had larger negative values and more consecutive negative months. The 1920s and 1940s show extended drought conditions. The 1980s also show dry conditions. The mid to late 1990s have a larger surplus of moisture than any of the other months in the record, but are unique amongst the predominately dry conditions that started in the early 1980s. July through October 1998 figures have negative indices although the months on either side are positive. This shows the strength of the evaporation over these months because the memory of drought in the index is strong, meaning if the previous years were wet it would have taken extremely dry conditions (such as those in 1998) to overcome them.
Looking more closely at the PDI index for the PIM, a strong moisture surplus is apparent before October 1999 and a tendency toward moisture deficit is apparent after this month (Figure 4-15). Many components of the analysis in this study have indicated the 1996-1997 and 1998-1999 water years are wetter than the average over the PIM. Larger inflow estimates, greater precipitation volumes, and lower evaporation rates occur before October 1999 compared with the rest of the period, with the exception of 1997-1998 when evaporation in the dry season was high. After October 1999, precipitation amounts are somewhat lower and evapo-transpiration rates are consistent or elevated with increased air temperatures as the open-water evaporation from the lake was shown to be. Therefore, resulting inflows were less.
Figure 4-15 - The Palmer Drought Index computed over the 1996-2005 period only based on air temperatures measured at the Sooke Dam.

Based on the evaluation of the different dry and wet seasons in the PIM, the dry season of 1998 and the wet season of 2000-2001 are the best examples of a dry season with extremely high evaporation and a wet season with extremely low water inputs from the PIM. The long-term record of the PDI has shown that these seasons are within the range of drought conditions that have occurred in the SR in the 1919 to 2005 record. In fact, drought conditions have been greater in magnitude and more persistent than they were in the 1998 dry season or the (2000-2001) wet season. In addition, conditions in the 1998 dry season were warmer than any other in the PIM and the mean annual temperature anomaly for the 1971-2000 base period was the largest at 1.4°C more than average in 1998. Mean annual air temperature is projected to be 2.6°C to 2.8°C warmer in this region in the 2050s (2041-2070) (Rodenhuis et al., 2007). Hence, because the dry season of 1998 is one of the warmest, it best
represents conditions that may occur in the SR in the future even though they are projected to be warmer.

To create a worst-case drought scenario, the inflows, precipitation, and evaporation from the 1998 dry season and the inflows, precipitation, and evaporation from the 2000-2001 wet season were used as representative values. Consumption rates for the dry season of 2001 were selected to illustrate the lowest rate at which supply was consumed over the record. Errors in inflows and change in storage were likely minimal in these dry and wet seasons because water was not spilled from the reservoir and water-level measures were taken hourly rather than daily to compute the storage. Other influences on the accuracy of these scenarios include the selected water restrictions and the assumed population. If water restrictions were reduced to Stage 2 or even Stage 1, it would have taken less time for the SR to reach critically low levels. In addition, these scenarios are based on the consumption rates of the current population. Population is expected to increase 26% by 2026 in the Greater Victoria Area, which would increase rates of consumption in spite of water restrictions (CRD, 2004b).

The initial starting conditions were set at 184.9 masl and then 186.0 masl to represent the average and maximum values for April 1st from 2003 to 2005 (the period over which the SR was at its maximum capacity). The change in storage for each consecutive season was then added to the initial condition, starting with the dry season first, followed by the wet season. For the average starting condition of 184.91 masl, if extreme conditions persisted it would take until some time in the third dry season for the water-level to go below 170 masl if water restrictions were not in place. Similarly, if the starting condition were at the maximum of 186.0 masl, water-levels would not go below 170 masl until late in the third dry season. However, if the rate of consumption over the summer months were equal to the average consumption rate for the dry season for the PIM, supplies would not fall below a useable volume until some time in the fifth dry season given average starting water levels or until late in the fifth dry season given maximum starting water levels.
In upgrading the dam to its current capacity, the CRD has safeguarded the population from the affects of drought. These drought scenarios have shown that this system is resilient. However, these scenarios assume that consumption rates are reduced to those accompanying level 3 water restrictions. Operating at this level for multiple years would not be practical as it would require the water users to forgo watering their lawns, washing their cars, etc. Hence, the potential resilience of the system under other demand side management measures, such as level 2 water restrictions, should be conducted. Furthermore, these scenarios were based from starting water levels that were average or above average. Under lower waterlevels the resilience of the system would be reduced.

Potential challenges in the face of drought could arise in the future with increased population and increased users of the water supply system. Additionally, the impact of climate change on maintaining optimum water quality has not been investigated here and efforts should be made in this regard as consequences could be large, especially with the anticipated increase in extreme events. Indices should also be developed to assist managers in deciding on when to put such water restrictions in place that are based on forecasting of PDO and ENSO events.

5 CONCLUSIONS AND FUTURE RECOMMENDATIONS

This was the first full account of the water balance for the SR, with the exception of groundwater, documenting the seasonality of the water balance and its extreme conditions. Exploration of the seasonal contrasts in the water balance demonstrated that they were strong. The wet season often brought more water to the supply than there was capacity to store, and strong evaporation and minimal precipitation over the dry season contributed significantly to moisture deficits. Through solving the water balance and linking it to data that have been measured over the long term (temperature and precipitation), the foundation for future studies of the influence of climate variability and change on the water balance of the SR has been set.
Two methods of estimating inflows were tested and each was shown to have strengths and weaknesses. Estimation of inflows demonstrated the importance of peak flows in this watershed to filling the reservoir. Significant volumes of water were lost if peak-flows were not accounted for by the modelling process applied. Leakage from the Butchart Reservoir via Maple Creek, presented challenges to understanding the linkage between climatic conditions and low-flows, and were difficult to model.

Losses from the SR include evaporation, spill, and consumption. Evaporation took place during the cooler wet season, but was minimal in comparison to inputs from precipitation and runoff. However, in all of the dry seasons except one, evaporation exceeded inputs from precipitation resulting in negative P-E values. This reflects not only a deficit from the SR, but also a withdrawal of moisture from the landscape, which effectively reduces the amount of runoff from the landscape that would result for the same amount of precipitation. On average, evaporation accounted for 9% of the water lost from the SR during the dry season. Spill was a major loss over the wet season in the earlier part of the PIM. Conversely, under the current capacity, there is little requirement for spill except to maintain water quality. Demand-side management practices had a large impact on rates of consumption.

Change in storage followed a distinct, re-occurring seasonal pattern dictated by inputs during the wet season that increased storage and continuous withdrawals that reduced storage by the dry season. Over the study period, there were a number of extreme events: two periods of extreme low-water conditions and one extreme high-water event. In the first low-water period, the 1997-1998 water year, April, May and June had low precipitation, and August had close to none. In the second period, the 2000-2001 water year, precipitation over the entire wet season was below normal. The 2003-2004 water year was marked by record-high rainfall volumes in October 2003.

Results of the Palmer Drought Index for 1919 to 2005, based on air temperature and precipitation, verified that the early half of the PIM had wetter conditions and the later half
was predominately dry. The low-water conditions in the PIM were comparable to other droughts that occurred in the 1920s and 1940s, but the durations of the droughts in the early part of the century were longer. This suggests that dry conditions have occurred here in the past. Therefore, it is possible that they could occur in the future for extended periods. Encapsulating extreme events in the PIM and using the PDI to verify that droughts have occurred here in the past lent credibility to the worst-case drought scenarios developed in this work.

Under persistent drought conditions with strict water restrictions, the drought scenarios demonstrated that water supply would be adequate for the Greater Victoria population until sometime in the fifth dry season, starting from average initial water levels. Supplies would last into late in the fifth dry season if initial conditions were set at the highest recorded water level for April 1st. The resilience of the system is limited to only three water years under drought conditions if average consumption rates are maintained and water levels start at either average or maximum conditions as recorded during the PIM. This demonstrates the resilience of the water supply when Stage 3 water restrictions are imposed. Some caution should be taken with these scenarios because they have investigated conditions when starting water levels in the SR were relatively high and have applied the highest water-restriction level, which would not be sustainable over the long term. Considering the projected pressures of population growth, climate change and variability on this supply the resilience of the system could become reduced in the future.

Improvements to the methods could be made through better measurement of discharge and upgrades to the HBV-EC model and its application. A more thorough temporal and spatial measurement of runoff over the catchment area could be made by selecting more stable cross sections and by refining the rating curves over multiple sites and years. Special attention should be paid to investigating the role of Maple Creek discharge in the dry period. It would be desirable to recalibrate the HBV-EC model with special attention to high flows.
Also, better representing physical parameters like variable interception rates with storm size could also improve the representation of observed flows. Further improvements to the estimate of inflows could be made by augmenting the numerical representation of routing in the HBV-EC model to handle low and high flows comparably, by adding a third reservoir.

Extensive work could be done to build an understanding of groundwater inflows and outflows, and how they contribute to the water balance, such as isotope mass balance. In addition, some of the techniques used to estimate precipitation and evaporation could be improved. There are five other precipitation stations in the catchment area, in addition to that at Sooke Dam, which could be used to characterize the gradients of precipitation across the basin by season. Once established, these gradients could be related to the long-term record at Sooke Dam. The estimate of evaporation could be improved by measuring wind speed and humidity at the ideal height of 2 m from the surface of the water. Finally, diversions into the reservoir from the Council Reservoir could be further assessed to better understand the possible error bounds on estimates.
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Vancouver Sun, November 7th, 2006. “It's a rainfall record for the province”


CHAPTER 5: SUMMARY

Population growth and climate change have put increased pressure on the water supply in recent years. These pressures are likely to increase in the future and threaten the reliable provision of water. Reservoir expansion options are limited by the lack of adequate sites near urban centres and by public resistance to the building of reservoirs that disrupt and degrade ecosystems. The easily accessed surface water stored in lakes or rivers is a small component of all the water on earth.

Providing a reliable water supply requires water conservation programs to be intensified and management strategies to become more efficient to meet these pressures. Strong planning efforts rest upon a sound understanding of the physical system under management. In the case of a water supply, dependable approaches should be built upon a thorough understanding of the individual components of the water balance at the highest level of precision possible. The hydroclimatology of a region is a pivotal driver of the water balance. It depicts the average climatic and hydrological characteristics unique to each region and is affected by climate change and variability. Hence, the hydroclimatology depicts the current behaviour and impacts the future response of a system.

The SR is situated in one of the most complex hydroclimatic regions in Canada where there is high inter- and intra-annual variability. This region is influenced by long- and short-term modes of climate variability, such as the multi-decadal PDO and the intra-decadal ENSO, because of its proximity to the Pacific Ocean. Climate change will likely increase air temperature, alter precipitation patterns, and increase the likelihood of extreme weather events in this region.

Data collection has taken place for nearly 100 years in the SRC, which provides an opportunity to test the influence of long-term climate change and variability. However, multiple expansions of the SR and alterations in the data collection procedures prompted
more in-depth analysis to better understand the system. Intense monitoring of streamflow, air temperature, water temperature, relative humidity, wind speed, and net radiation took place from October 1996 to September 2005 provided an opportunity to quantify the water balance of the SR.

Over the PIM, there was a wide range of hydroclimatic conditions representative of those that occurred in the long-term record. From October 1996 to March 1998, the climate was slightly wetter than average while from April 1998 to September 2005 it became much drier. The lowest annual rainfall occurred in 2000; the highest occurred in 2003. Three of the hottest summers since 1916 occurred in 1998, 2003, and 2005 respectively. The 1998 air temperature was the highest, but was not quite as high as those predicted for the 2020s for the region with multiple GCMs.

In spite of this intensive monitoring during the PIM, a full depiction and independent estimate of the inflows into and the evaporation from the SR had not been completed. The high air temperatures and limited precipitation in the dry season indicated the importance of evaporation to the water balance. They also showed the strong intra-annual variability or seasonality of the SR. Evaporation was investigated by three different methods. The objectives included (a) defining the influence of heat storage on evaporation; (b) estimating evaporation over the PIM; (c) exploring the strengths and limitations of the applied estimates with the given dataset; and (d) testing a temperature-based estimate for use during periods when climate observations were limited, such as the period from 1916.

Evaporation was estimated using the mass-transfer and radiation-based Penman method, radiation-based Priestley-Taylor method, and the temperature-based Hamon method. Results showed heat storage to be an important influence on the timing and magnitude of evaporation. Heat stored in the lake contributed to evaporation peaking a month later and increasing in intensity by up to 53 mm per month, compared with that estimated without heat storage. Heat was released from the SR throughout the fall and into the early winter,
providing the energy and vapour-pressure gradients that drove evaporation in the low-radiation periods.

Evaporation rates over the PIM were found to have a strong seasonality with high values over the dry season; low yet substantial values over the wet season; and high annual amounts relative to other sites in North America. The Penman and Priestley-Taylor estimates included heat storage and the Hamon did not. The Penman approach was thought to produce the most reliable estimate because it was shown in other studies to copy the energy-balance method estimates most closely. Though wind data was not collected at the ideal location for use of the Penman method, testing against the Priestley-Taylor method (which does not use wind speed) showed that wind-speed measurements did not inhibit the accuracy of the Penman. The values from the Priestley-Taylor method were found to over-estimate the evaporation slightly, as has been found in other studies. The Hamon method had low annual estimates compared with the other techniques, and did not show the same monthly or seasonal trends.

A technique was developed to modify the Hamon estimate to copy the Penman estimate, to allow the estimation of evaporation when climate data was limited. This method was used to adjust the monthly Hamon values to reflect the influence of heat storage on the seasonality of evaporation and to address the insufficiency of the Hamon method to derive reasonable annual totals for this location. This was done by dividing the monthly average Hamon estimates by the Penman estimates. This method had some success; seven of the twelve months produced estimates that were close to Penman estimates, but five of the twelve did not. Estimates of evaporation for June and August were the least similar to the Penman estimates because of the high inter-annual variation in net radiation in these months. The months of February, March, April, May and July were estimated with the most accuracy, probably because they had the lowest inter-annual fluctuation in net radiation.
After the completion of the dam in December 2002, heat-storage patterns started to differ from those seen previously, because the increased volumes and altered withdrawal patterns affected the storage and release of heat in the SR. The dam had been modified twice in the past, which likely altered the heat storage responses with each new regime. Because of this, the heat storage correction may not provide reliable estimates of evaporation before the dam was modified in 1970. Additionally, this relationship is best suited to the climatic conditions for which it was developed and may not work as well if the setting changes due to climate variability or change. Hence, future work should test a multiple linear regression equation that could explain the relationship between air temperature and heat storage, net radiation, and vapour pressure deficit, to develop a modified Hamon estimate that would produce more reliable estimates under changing-dam capacity.

The water balance total inflows into the SR were defined. Two methods were used to address total inflows:

1. A contributing area approach where inflows from the monitored catchments were scaled up to account for runoff from the whole catchment area; and
2. A hydrologic model was calibrated to one watershed and then applied to the other watersheds, and the volumes from these watersheds were scaled up to estimate runoff from the whole catchment area.

Both methods had limitations. The contributing area approach tended to over-estimate low flows and the hydrologic modelling technique tended to under-estimate high flows. It was found that total inflows were best estimated with the contributing area approach, because the under-estimate of peaks during high runoff events by the modelling technique significantly underestimated total inflows when multiplied across the entire SRC and summed over a given month. Hence, the contributing area method was used in the water balance.

With inflows and evaporation estimated, precipitation studied previously, and storage, consumption, spill, and fisheries releases measured by the CRD, the water balance
was evaluated to estimate change in storage. This \textit{calculated} change in storage was compared to the \textit{observed} change in storage. Closure of the water balance was tight in the dry season and during the wet season when spill did not take place. Spill measurements before December 1998 were found to contribute significant uncertainty to the water balance.

The water balance was investigated on a monthly basis and then seasonally over the wet (October-March) period and the dry (April-September) period. The largest input to the system is runoff during the wet season. The greatest storage increases usually occurred during the months of February and March variability from year to year was not great. The largest losses were during August, September, and October, when inputs were low and consumption and evaporation were high. In seven out of nine years, the SR filled and extra water had to be spilled. In some water years, the SR was full at the end of the wet season but did not have sufficient capacity to provide water for the following dry seasons, when high evaporation rates led to large water losses. This occurred before the capacity of the SR had been increased. Evaporation estimates put into context compared with the relative number of other water-balance components, such as precipitation and inflows, showed that evaporation was often extensive, compared with inflows and precipitation in the dry season. The monthly and seasonal inputs of precipitation and inflows and outputs of evaporation and consumption varied greatly from year to year.

The SR is managed at only 96% reliability, which means that statistically the SR is managed so that it can reach extreme lows once every 25 years. In two of the nine years in the PIM, storage in the SR reached extremely low levels when strict demand side-management practices had to be implemented to sustain the supply. Studies have shown that with climate change comes an increased probability of extreme events, and an average temperature increase from 2° C to 5° C. Under these conditions, high evaporation rates will become more likely. Projections of precipitation are not consistent in their magnitude and direction.
In the future, it is possible that conditions could become both wetter and drier; precipitation events will increase in magnitude in the wet season and the dry season will have less total rainfall and persist over a longer period. This combination will exacerbate the challenges of a supply where storage is sometimes insufficient to capture the precipitation falling in the wet season, and the dry season has limited precipitation and high evaporation rates. Planning will become more challenging in the future for the SR because the water-balance components already have high inter- and intra-annual variability.

The information gathered from the water balance over the PIM was used to test the resilience of the SR during times of persistent low inputs from precipitation and runoff and high loss from evaporation. Water consumption rates were set at their lowest on record where strict demand side management practices were in place. It was found that, under these conditions, the SR could provide supply into the fifth dry season if water levels on April 1st were equal to the average or maximum capacity measured on April 1st during the PIM for the current dam capacity. However, if strict water restrictions were not put in place, supply would last only until the third dry season.

This work included careful analysis of all terms of the water balance equations for the SR, which revealed some important findings. Closure of the water balance was hampered at periods in the record that coincided with spill events and in many cases the amount of water that could not be accounted for was proportional to spill amounts. This suggests that the estimation of spill had some errors, especially prior to 1998. Since the capacity of the dam has increased it has not been necessary to spill as often. However, there is still potentially error associated with these measurements. Inflows estimates also contributed a large relative proportion of error. The quick response time of this system made it challenging to maintain reliable stage-discharge curves and to model. Annually, errors in precipitation and evaporation are proportionally small. However, due to the large role of evaporation in the dry season, errors in estimates at the monthly or seasonal scale are important.
Future work towards refining the analysis of the water balance should focus on re-evaluating rating-curve relationships, taking a new approach with hydrologic modelling and installing more measurement devices. Checking spill-water-level relationships to ensure that all future estimates are within narrow ranges of error should be a priority. After that, inflows estimates could be improved with review of rating curves, gauging of more catchments for multiple seasons to better understand their relationship to Rithet and Judge Creeks, and further model calibration to better match peak-flow events. Additionally, Maple Creek should be monitored and discharge should be related to the pressure head created by Butchart Reservoir to tie its flows to the longer term record off water-level for this reservoir.

Two of the measurements devices which collect data for estimating evaporation, relative humidity and wind speed, are not measured at suitable locations. Ideally, they would be measured 2 m above the surface of the lake. These instruments are at unfavourable heights and also near turbulent boundaries, which alter air flow and potentially impact wind speeds and humidity levels. Evaporation estimates would also benefit from more regular measurement of heat storage in the reservoir. This would allow the thermal budget periods, over which meteorological parameters are averaged in estimating evaporation, to be more uniform and cross-comparable. Furthermore, stable isotopes tracers could be employed to determine ratios of inflow to evaporation and employed as a check on evaporation estimates.

Finally, the relationship of air temperature, precipitation, and evaporation to synoptic climatology should be established. This information would provide a link between the water balance of the SR and climate variability and change. Synoptic weather patterns can be projected more reliability with Global Climate Models than climate at a point. Thus, linking the water balance to these patterns in the past may help to identify how the SR might respond to climate change in the future.
Appendix

Appendix A
Figure a. Stage-Discharge Relationship and Hypsometric Curve for Horton Creek.

\[ Q = 4.759 \times (h_1 - 0.196)^{2.821} \]
RMSE = 0.18 m³ s⁻¹

Figure b. Stage-Discharge Relationship and Hypsometric Curve for Whiskey Creek.

\[ Q = 6.722 \times (h_1 - 0.080)^{2.302} \]
RMSE = 0.25 m³ s⁻¹

Figure c. Stage-Discharge Relationship and Hypsometric Curve for Maple Creek.
Appendix B

Figure a. Regression of Horton Creek and Maple Creek to Rithet Creek Discharge.

\[ y = 6 \times 10^{-15}x^3 - 5 \times 10^{-8}x^2 + 0.1948x + 99740 \]
\[ R^2 = 0.9843 \]

\[ y = 8 \times 10^{-9}x^2 + 0.0783x \]
\[ R^2 = 0.9785 \]

Figure b. Regression of Whiskey Creek to Judge Creek Discharge.

\[ y = -8 \times 10^{-15}x^3 + 3 \times 10^{-8}x^2 + 0.3745x \]
\[ R^2 = 0.9968 \]

\[ y = 8 \times 10^{-9}x^2 + 0.0783x \]
\[ R^2 = 0.9785 \]
List of Symbols

\( I_s \) = surface inflows (m³ month⁻¹),
\( I_g \) = inflows from groundwater (m³ month⁻¹),
\( P \) = precipitation onto the reservoir (m³ month⁻¹),
\( E \) = evaporation from the reservoir (m³ month⁻¹),
\( O_s \) = surface outflows (m³ month⁻¹),
\( O_g \) = groundwater outflows (m³ month⁻¹),
\( \Delta S \) = change in storage (m³ month⁻¹),
\( O_{s1} \) = surface outflows via consumption (m³ month⁻¹),
\( O_{s2} \) = surface outflows via spill (m³ month⁻¹),
\( O_{s3} \) = surface outflows via fisheries release (m³ month⁻¹),
\( \varepsilon \) = the error accumulated from each term (m³ month⁻¹),
\( \varphi \) = the measure of closure of the water balance (%),
\( E_p \) = Penman evaporation (mm month⁻¹),
\( E_{pr} \) = Priestley–Taylor evaporation (mm month⁻¹),
\( \nu = 1.26 \) = Priestley–Taylor empirically derived constant (dimensionless),
\( s \) = slope of the saturated vapour pressure–temperature curve at mean air temperature (Pa °C⁻¹),
\( \gamma \) = psychrometric “constant” (depends on temperature and atmospheric pressure) (Pa °C⁻¹),
\( R_n \) = net radiation (MJ m⁻² d⁻¹),
\( R_{ns} \) = net short wave radiation (MJ m⁻² d⁻¹),
\( R_{nl} \) = net long wave radiation (MJ m⁻² d⁻¹),
\( \alpha \) = the reflection coefficient or albedo (dimensionless),
\( Q_{hs} \) = change in heat stored in the water body (W m⁻²),
\( \Delta T_{wi}(z) \) = the change in water temperature between the measurement on the second day (°C) and the first day at depth \( z \) (m),
\( \Delta t \) = the number of days between measuring intervals (converted to seconds),
\( a(z) \) = the lake area at depth \( z \) (m),
\( \Delta z \) = the layer thickness (m; typically 1 m),
\( f_w \) = wind function (m s⁻¹),
\( a_a \) and \( b_a \) = wind function coefficients (dimensionless),

\( U_h \) = the wind speed (m s\(^{-1}\)) at height \( h \) (m),

\( U_2 \) = the adjusted wind speed (m s\(^{-1}\)),

\( \lambda \) = the latent heat of vaporization (MJ kg\(^{-1}\)),

\( D = (e_s - e_a) \) is the vapor pressure deficit (kPa),

\( e_s = \) saturation vapour pressure (kPa),

\( e_a = \) actual vapour pressure (kPa),

\( e_{S(\text{avg})} = \) mean saturation vapor pressure (kPa),

\( RHI = \) relative humidity (%),

\( \rho = \) density of water (998 kg m\(^{-3}\) at 20 °C),

\( U_2 = \) adjusted windspeed at 2 m above surface (m s\(^{-1}\)),

\( \Pi = \) the atmospheric pressure (kPa),

\( T = \) air temperature (°C),

\( T_t = \) the daily total atmospheric transmittance (dimensionless),

\( \Delta T = \) the daily range of air temperature (°C),

\( A, B, \) and \( C \) are empirical coefficients (dimensionless), determined for a particular location from measured solar radiation data,

\( Q_o = \) the daily extraterrestrial insolation incident on a horizontal surface (J m\(^{-2}\)),

\( S_o = \) the solar constant (1360 W m\(^{-2}\)),

\( \bar{d} = \) the mean value of the distance from sun to earth (m),

\( d = \) the distance from sun to earth (m), \( \left( \frac{\bar{d}}{d} \right)^2 \) was taken as unity because it does not differ from unity by more than 3.5%,

\( h_s = \) the half day length (cos \( h_s = -\tan \phi \tan \delta \) radians),

\( \phi = \) the latitude of the location of interest (radians),

\( \delta = \) the solar declination (radians),

\( f = \) adjustment for cloud cover (dimensionless),

\( \epsilon' = \) net emissivity between the atmosphere and the ground (units?),

\( \sigma = \) Stephan-Boltzman constant = 4.903 x 10\(^{-9}\) (MJ m\(^{-2}\) °K\(^{-4}\) d\(^{-1}\)),

\( R_{SO} = \) clear sky radiation (MJ m\(^{-2}\) d\(^{-1}\)) with values for average climates,

\( R_A = \) radiation (MJ m\(^{-2}\) d\(^{-1}\)),

\( d_r = \) the relative distance between the earth and the sun (m),

\( J = \) Julian date,

\( E_H = \) Hamon evaporation (mm month\(^{-1}\)).
$L$ = the day length (hours),
$\text{PIM} = \text{Period of Intensive Monitoring}$,
$\psi_m = \text{the monthly conversion factor (dimensionless), where } m \text{ is the month},$
$Q = \text{discharge (m}^3\text{ s}^{-1}\text{)},$
$Q' = \text{predicted discharge (m}^3\text{ s}^{-1}\text{)},$
$\varpi = \text{coefficient (dimensionless)},$
$\beta = \text{coefficient (dimensionless)},$
$h = \text{stage (m)},$
$h_o = \text{a zero stage reference datum (m), } \alpha \text{ and } \beta \text{ are dimensionless coefficients},$
$Q_1 = \text{discharge from the modelled area (m}^3\text{ s}^{-1}\text{)},$
$A_1 = \text{area that is modelled (m}^2\text{)},$
$A_2 = \text{area of the representative area (m}^2\text{)},$
$Q_2 = \text{total discharge from the representative area (m}^3\text{ s}^{-1}\text{)},$
$I_\eta = \text{total inflows estimated by HBV-EC modelling approach (m}^3\text{)},$
$I_X = \text{total inflow volumes estimated with contributing area approach (m}^3\text{)}.\]