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Historic and Likely Future Impacts of Climate Change on Lake Tahoe, California-Nevada, USA

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14.1 Introduction and background

Lake Tahoe is a large ultra-oligotrophic lake lying at an elevation of 1898 m in the central Sierra Nevada on the California–Nevada border (Figure 14.1). The lake is renowned for its deep cobalt blue color and clarity. Due to concerns about progressive eutrophication and loss of clarity, the lake has been studied intensively since the mid-1960s, and has been the focus of major efforts to halt the trends in clarity and trophic status (Goldman 1981; TERC 2011). Previous work on the effects of climate change on the lake (Coats *et al*. 2006) showed (i) that the lake is warming at an average rate of about 0.013 °C year^{-1}; (ii) the warming trend in the lake is driven primarily by increasing air temperature, and secondarily by increased downward long-wave radiation; (iii) the warming trend on monthly and annual timescales is correlated with the Pacific Decadal Oscillation (PDO) and (to a lesser extent) with El Niño-Southern

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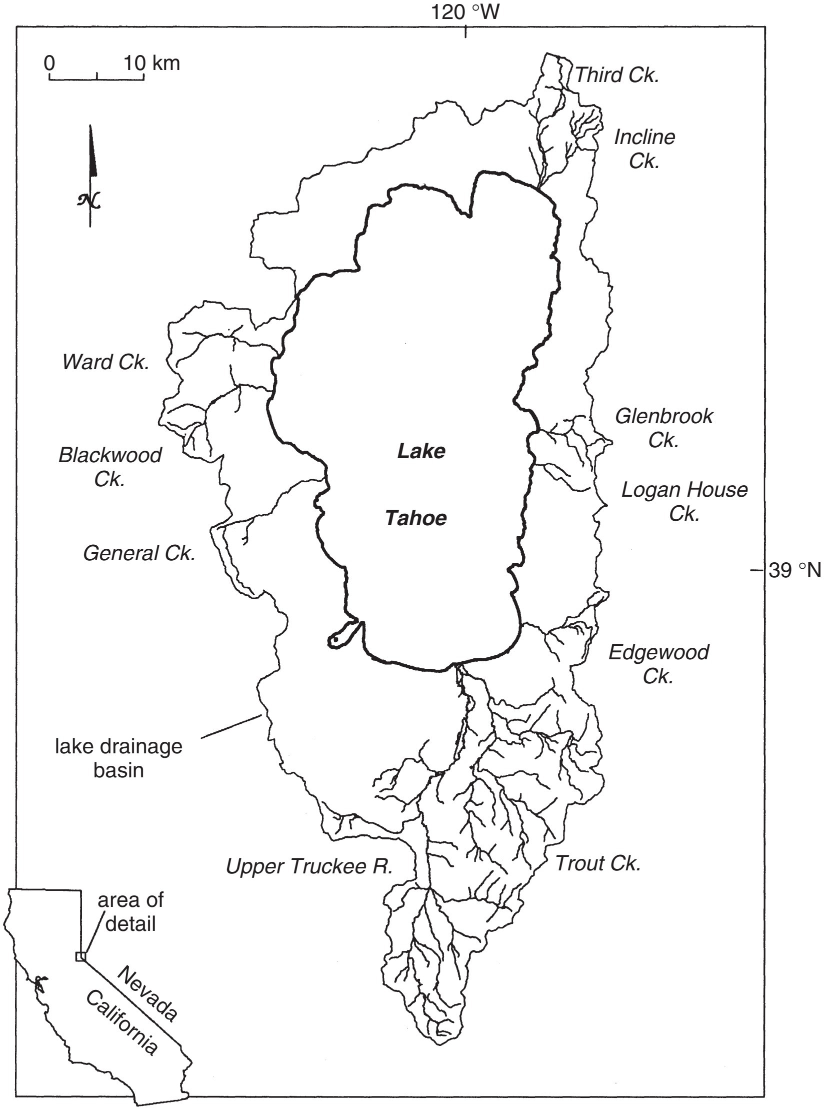


Figure 14.1 Map of the Tahoe Basin.

Oscillation (ENSO); and (iv) the warming of the lake is modifying its thermal structure, and increasing its resistance to deep mixing. Table 14.1 summarizes additional information about Lake Tahoe.

Warming trends and changes in thermal structure have been identified in lakes of Europe, Africa, and North America. Analysis of a 52-year record of monthly temperature profiles in Lake Zurich showed a secular temperature increase at all depths, resulting in a 20% increase in thermal stability. A temperature model of that lake showed that the warming trend in the lake is most likely explained by increasing night-time air temperature, concomitant with reduced night-time rates of latent and

Table 14.1 Basic facts about Lake Tahoe.

Volume	156 km^3
Surface area	498 km^2
Average depth	313 m
Maximum depth	505 m
Depth of photic zone	To ca. 100 m
Basin area	1310 km^2
Maximum lake elevation	1899.54 m NAVD
Mean hydraulic residence time	ca. 650 years
Mean annual lake discharge	2×10^8 m^3
Mixing frequency below 450 m	ca. 1 year in 4
Rate of decline in Secchi depth	ca. 25 cm $year^{-1}$
Limiting nutrient	N until early 1980s; now P

sensible heat loss from the lake surface (Peeters *et al*. 2002; Livingstone 2003). Warming of Lake Tanganyika between 1913 and 2000, associated with increasing air temperatures, has increased the vertical density gradient and thus decreased both the depth of oxygen penetration and the nutrient supply in the upper mixed layer (Verburg, Hecky, and Kling 2003).

In North America, modeling and statistical analysis at Lake Mendota, Wisconsin, showed that increasing air temperature is related to higher epilimnion temperatures, earlier and more persistent thermal stratification, and decreasing thermocline depths in late summer and fall (Robertson and Ragotzkie 1990). Twenty years of temperature records from the Experimental Lakes Area in Northwestern Ontario showed an increase in both air and lake temperatures of 2 °C, and an increase in length of the ice-free season by three weeks (Schindler *et al*. 1996). Modeling studies of the effects of climate warming on the temperature regime of boreal and temperate-zone lakes are consistent with these observations, predicting increased summer stratification, greater temperature increases in the epilimnion than in the hypolimnion, and increased length of the ice-free season (Elo *et al*. 1998; Stefan, Fang, and Hondzo, 1998). From 1979 to 2006, Lake Superior warmed at an average rate of 0.11 °C $year^{-1}$, about twice the rate of regional air temperature, due in part to a positive feedback from reduced ice cover and reduced albedo (Austin and Colman 2007). At Lake Michigan, application of a one-dimensional temperature model showed that climate warming will decrease the summer thermocline depth, increase resistance to mixing, and could lead to permanent stratification (McCormick and Fahnenstiel 1999). Closer to Lake Tahoe, Arhonditsis *et al*. (2004) found a warming trend in Lake Washington, and showed that the relationship between lake temperature and climate may be different during lake cooling than during lake warming, though the ENSO and PDO indices were positively correlated with lake temperature during both the warming and cooling phases.

Recently, the trends in night-time summer surface temperatures of six lakes in northern California and Nevada—Tahoe, Almanor, Clear, Pyramid, Walker and Mono—were measured using ATSR and MODIS satellite sensor data (Schneider *et al*. 2009). The upward temperature trend (1992–2008) averaged 0.11 °C $year^{-1}$, about twice the trend in surface air temperatures, though none of the lakes freeze in winter. The trend of temperature rise in the surface waters of Lake Tahoe was 0.13 °C $year^{-1}$ (1970–2007), ten times the rate for the entire lake volume.

14.2 Historic climate trends in the Tahoe Basin

Previous studies of the historic trends in hydroclimatology in the basin and nearby locations (1910–2007) indicated strong upward trends in air temperature (especially night-time temperature), a shift from snowfall to rain, a shift in snowmelt timing to earlier dates, increased rainfall intensity, and increased interannual variability (Coats 2010). The strongest warming trends for the period 1956–2005 were in T_{min} at Tahoe City (on the northwest side of the Lake), especially in summer months. At Truckee and Boca (~20–25 km north of Tahoe City and outside the Tahoe basin), both T_{max} and T_{min} also trended upward in winter and spring.

As might be expected, the trend in air temperature in the Tahoe basin is reflected in both the form of precipitation, and the timing of snowmelt. Figure 14.2 shows the fraction of annual precipitation at Tahoe City that fell as snow (that is, precipitation on days with average temperature $<0\,^{\circ}$C) from 2010 to 2007. Since total annual snowfall affects the timing of snowmelt peak discharge, the timing of the latter was examined by testing the slope of the time trend in residuals after removal of the effect of total annual snowfall. On average, the timing of the spring snowmelt peak discharge (1961–2005) for five monitored streams in the Basin has shifted toward earlier dates at a rate of about 0.4 days year^{-1}. No such trend was observed for the streams outside of the basin. This was consistent with the observation of Johnson, Dozier, and Michaelsen (1999) from 30 years of snow survey data from 260 snow survey courses in the Sierra Nevada that the Tahoe basin experienced the highest loss—54%—in May snow-water equivalent (SWE) of any of the 21 river basins studied.

Good temperature profiles for the lake are available from late 1969 to the present. These records allow us to examine the time trends in temperature at a given depth, and in the volume-averaged temperature for the entire lake (Coats *et al*. 2006). Figure 14.3 shows the temperature at 400 m depth for the period 1969–2007. The great volume of

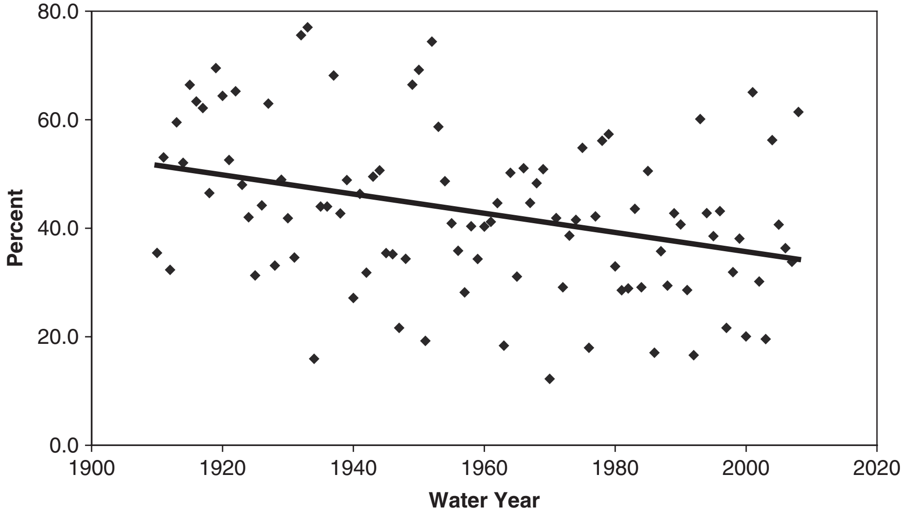


Figure 14.2 Percentage of annual precipitation falling as snow (defined as precipitation on days with mean temperature $<0\,^{\circ}$C).

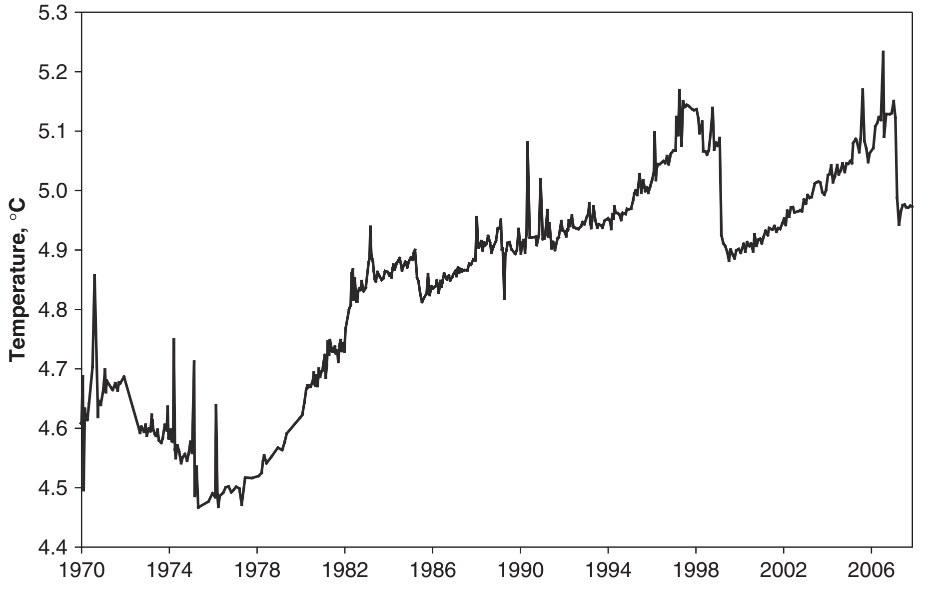


Figure 14.3 Temperature measured at 400 m in Lake Tahoe.

the lake filters out the short-term noise and reveals the long-term trend in temperature. Note the saw-tooth pattern, which is typical of warming temperate-region lakes (Livingstone 1997). During periods without deep mixing, heat is gradually transported downward by eddy diffusion. Then when the lake mixes to the bottom and becomes isothermal (typically during the late winter), the temperature drops sharply.

Figure 14.4 shows the trend in the deseasonalized volume-averaged lake temperature, 1970–2007. An interesting feature of this plot is the anomalous drop in temperature in 1982–1983. This was a period of a strong ENSO event, which is typically associated with warmer conditions. The short cooling trend was likely related to the March 1982 eruption of El Chichón in Chiapas, Mexico, which is known to have cooled air temperatures world-wide (Kerr 1982). The Seasonal Kendall Test (Salmi *et al*. 2002; Helsel and Frans 2006) for trend in monthly average lake temperature found a tau correlation coefficient of 0.54 and a Sen's slope of 0.013 °C year^{-1}, with $P < 5 \times 10^{-5}$.

The slope of the trend in volume-averaged temperature of Lake Tahoe is highest in September-November, and least in January-February, suggesting that increased summer warming plays a more important role that suppressed winter cooling (Coats *et al*. 2006). This contrasts with Lake Geneva, where Lemmin and Amouroux (Chapter 12) found that reduced winter cooling in the eplimnion and metalimnion played a larger role in the upward temperature trend than summer warming.

14.3 Modeling the impacts of future climate change

Understanding the historic and likely future conditions of Lake Tahoe's water quality and famed optical transparency requires consideration of the input of water, nutrients,

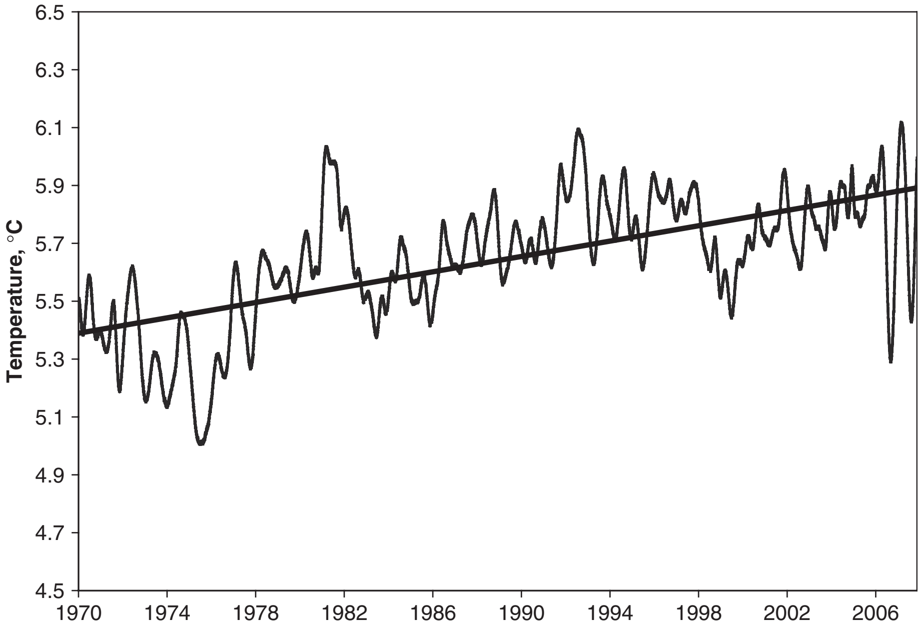


Figure 14.4 Volume-averaged daily temperature of Lake Tahoe, with seasonal effect removed.

sediment and energy from the lake's watershed and from the atmosphere (Jassby, Reuter and Goldman 2003; Reuter *et al*. 2003; Sahoo, Schladow and Reuter 2010). With continued change toward a warmer climate (IPCC 2007; Hansen *et al*. 2008), both research scientists and resource managers in the Tahoe basin would like to know: (i) How fast will the air temperature in the basin rise? (ii) How will the form, timing and annual amount of precipitation change? (iii) How will the changes in temperature and precipitation affect drought conditions, which in turn will affect vegetation, fire frequency and soil erosion? (iv) How will changes in precipitation affect streamflow regimes, especially high- and low-flow frequency-magnitude relationships? (v) How will continued warming of the lake affect its thermal stability, biogeochemical cycling and primary productivity?

Answering these questions is a daunting task, since the General Circulation Models (GCMs) used in global change research typically produce output at a 2.5° grid scale, too coarse to resolve the topographic complexity that drives climate in mountainous regions. Furthermore, to address extreme hydrologic events (precipitation and floods), an hourly time scale is needed, whereas GCM outputs are typically at daily or longer time scales.

To address these challenges, we used the twenty-first century downscaled daily temperature and precipitation (Hidalgo, Dettinger and Cayan, 2008; Dettinger, in press) from two GCMs (the Geophysical Fluid Dynamics Laboratory, or GFDL CM2.1 (Delworth *et al*. 2006) and the Parallel Climate Model, or PCM1 (Washington *et al*. 2000)) and two emissions scenarios (A2 and B1; Nakicenovic *et al*. 2000). For the GFDL, we also had wind (bias corrected), radiation and relative humidity data. The precipitation and wind data were corrected for bias by quantile mapping (Wood *et al*. 2001, 2002), to ensure that the empirical probability distribution functions matched the real station

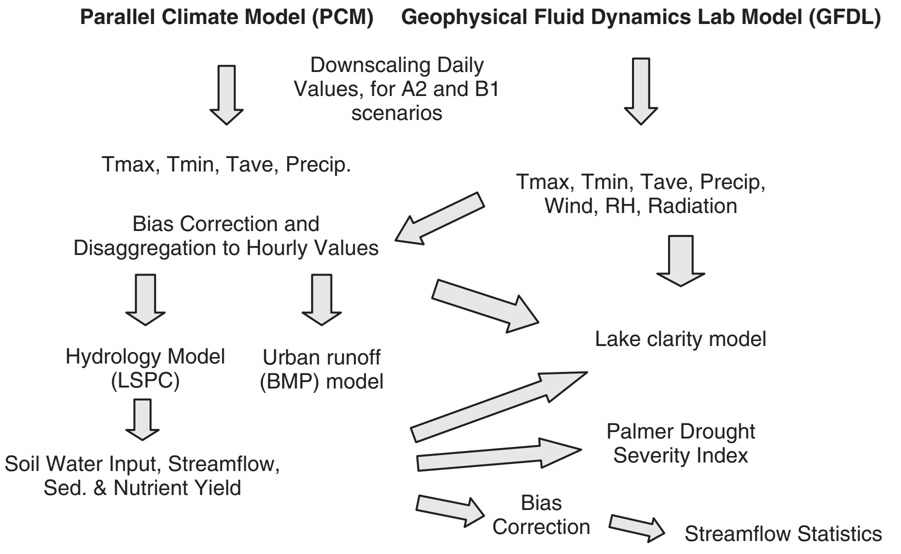


Figure 14.5 Information flowchart for project to model future impacts of climate change in the Tahoe basin.

data. The daily data were also adjusted using meteorological data for 12 SNOTEL stations (see http://www.wcc.nrcs.usda.gov/snotel/California/california.html, accessed July 11, 2012) in the Basin, and then disaggregated statistically to hourly values and used to drive a distributed hydrologic/water quality model—the Load Simulation Program in C++, or LSPC—for streams in the Tahoe basin (Riverson, *et al*. in press). The output from that model, along with the meteorological data, was used as input to a one-dimensional hydrodynamic/clarity model of the lake (Sahoo, Schladow and Reuter, 2010; Sahoo *et al*., in press), and (with additional bias correction) to calculate stream-flow statistics for the Upper Truckee River (UTR) including flood frequency and runoff timing. It was also used to calculate soil water input as the sum of rainfall and snowmelt, for calculation of the weekly Palmer Drought Severity Index (PDSI) for selected sites (Palmer 1965; Wells, Goddard and Hayes 2004). The hydrology data were used in a stormwater routing and water-quality model to investigate the implications of climate change for design of Best Management Practices (BMPs) in the Tahoe basin (Wolfe 2010). Figure 14.5 shows the information flow in the study design (see Costa-Cabral *et al*. in press).

14.4 Impacts on the watershed

14.4.1 Precipitation and drought

Climate modeling studies of likely future precipitation in northern California reach no consensus on the direction of trends, possibly because that region is near the transition between the mid- to high-latitude zone of increasing precipitation and a band of drying conditions over the subtropical north Pacific and Mexico (Dettinger *et al*. 2004). Figure 14.6 shows the projected trends in total annual precipitation averaged over the

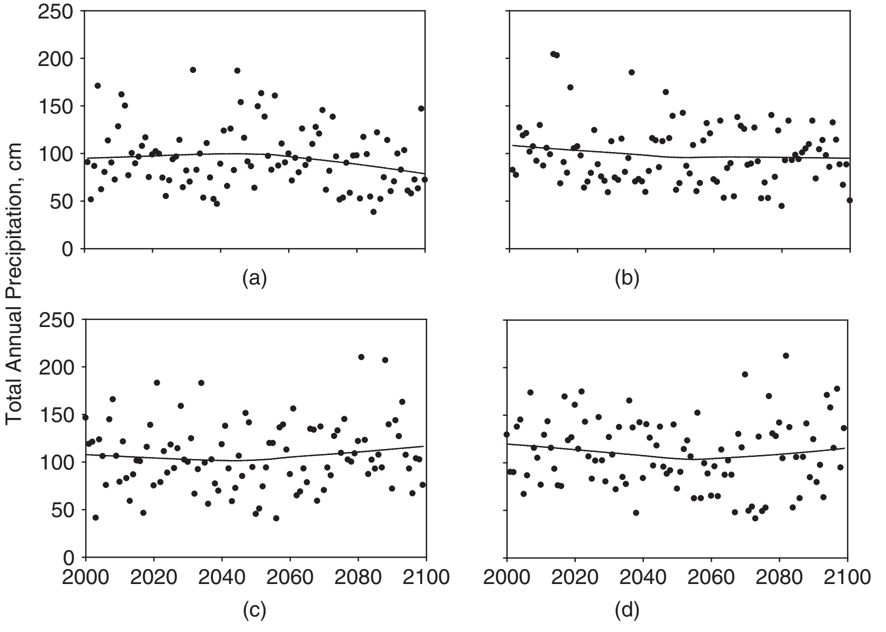


Figure 14.6 Total annual precipitation averaged over the Tahoe basin, for two models and two emissions scenarios. Curves are from a LOWESS smoothing (a) GFDL A2, (b) GFDL B1, (c) PCM A2 and (d) PCM B1.

Tahoe basin, from the two GCM models for each emissions scenario. For the GFDL A2 case, the downward trends in annual precipitation were significant ($P < 0.01$) for the months of October, December, April, May and June as well as for the annual total. For the GFDL B1 case, the downward trends were significant for January and December, as well as for the annual total. As shown in Figure 14.6, precipitation was projected (for the GFDL A2 case) to drop most sharply toward the end of the century. The 20-year running average for that case reached a maximum of 114 cm year^{-1} in 2060, and then dropped to a minimum of 79 cm year^{-1} in 2098. The PCM results, however, did not show major changes in future precipitation in the basin (see Coats *et al*. in press for more detail).

Changes in the form of precipitation, however, may be more important than changes in total amount. Figure 14.7 shows the expected trend in percent of precipitation falling as snow, averaged over the Basin, for the most contrasting scenario-model combinations. The loss of snowpack will have an effect on available soil moisture that is greater than the effect of precipitation change alone. Figure 14.8 shows the projected trends in the PDSI (expressed as the annual driest week), for the GFDL results at sites on the west (Tahoe City) and east (Glenbrook) sides of the basin. Note the sharp downturn for Glenbrook, where the annual snowpack is already thin and ephemeral in some years. At that site, the percent of annual soil water input as snowmelt explained 26% of the variance in the modeled maximum annual drought severity (minimum annual weekly PDSI value), with $P < 10^{-7}$.

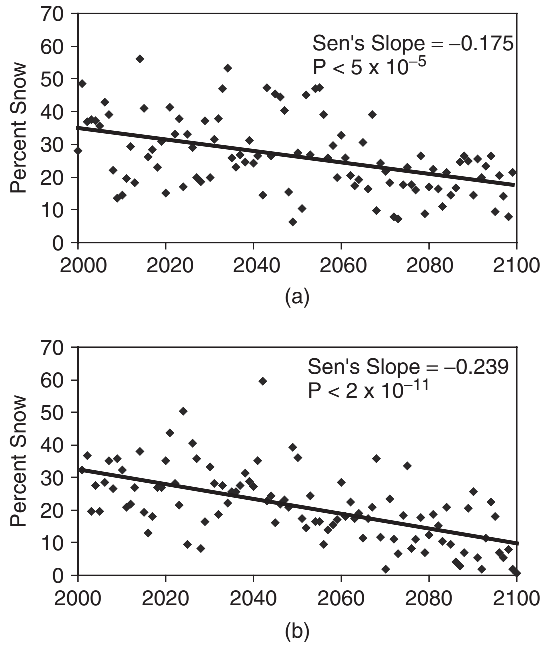


Figure 14.7 Percent of annual precipitation falling as snow averaged over the Tahoe basin. The B1 emissions scenario with the PCM (a) represents the least shift from snow to rain of the four cases modeled, and the A2 scenario with the GFDL model (b) represents the greatest shift.

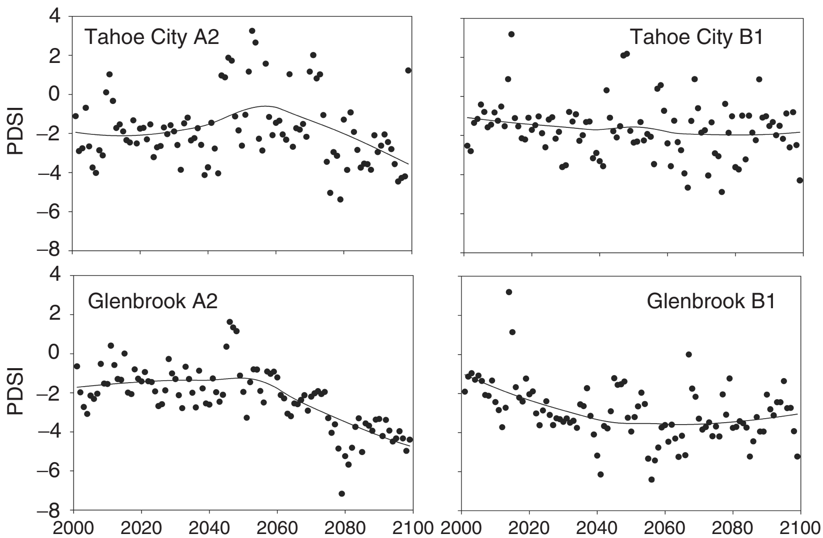


Figure 14.8 Trends in the modeled annual minimum weekly Palmer Drought Severity Index, for two locations (east and west sides of the Tahoe basin) and two emissions scenarios. Soil moisture input is calculated as daily rain plus snowmelt using a distributed hydrologic model for the basin.

14.4.2 Runoff

A number of studies have shown how climate change is shifting the timing of stream runoff in the western US, toward earlier dates, both historically (Lettenmaier and Gan 1990; Aguado *et al*. 1992; Wahl 1992; Pupacko 1993; Brown 2000; Stewart, Cayan and Dettinger 2005; Coats 2010) and in modeled future scenarios of climate change (Dettinger and Cayan 1995; Dettinger *et al*. 2004; Maurer 2007). Figure 14.9 shows the projected trends in the timing of the annual hydrograph centroid (also called the Center timing, or CT), for the Upper Truckee River, the lake's largest tributary. This measure takes account of the changes in precipitation form and amount throughout the water year, whereas the historic shift in snowmelt timing is more sensitive to warming trends in the spring. For the last third of the twenty-third century, the projected flow duration curve for the Upper Truckee shows a distinct downward shift for the A2 scenario (consistent with the decreasing precipitation and increasing drought), but not for the B1 scenario.

The projected changes in runoff timing point toward reduced summer low-flow in basin streams, at least for the A2 scenario, but the statistical analysis does not show the full effects. Most of the streams in the Tahoe basin streams flow through coarse alluvium in their lower reaches, so even a small drop in summer stream flow can cause a shift from surface to subsurface flow, and thus destroy aquatic habitat.

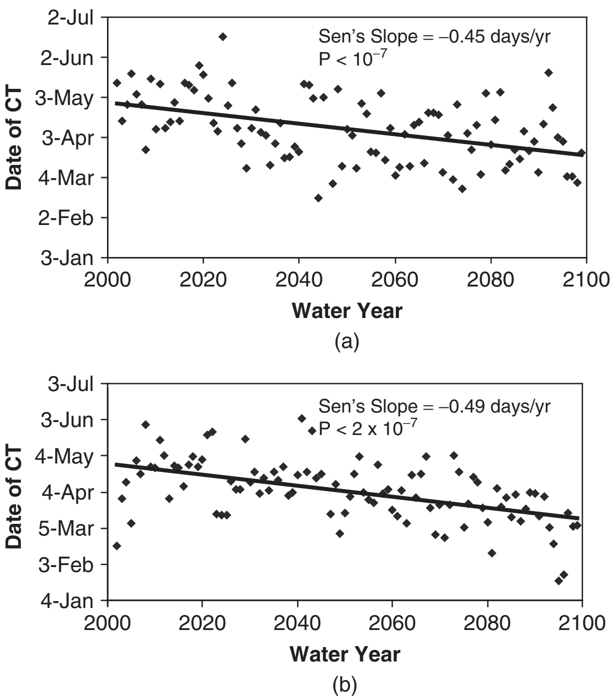


Figure 14.9 Modeled trends in the date of Center Timing of annual runoff, Upper Truckee River, for A1 and B2 scenarios, by GFDL (a) GFDL B1 and (b) GFDL A2.

14.4.3 Floods

Although climate change in the Tahoe basin will have a chronic effect on stream low-flow, the most dramatic effects may be on the flood frequency-magnitude relationship. Our future flood frequency curves created from downscaled GFDL/LSPC output were adjusted downward, using an equation that mapped the modeled GFDL flood frequency curves for the 1972–1999 period onto the actual measured curves for the same period. Figure 14.10 shows the modeled changes in flood frequency for the Upper Truckee River for three thirds of the twenty-first century compared with the historic curves from USGS data (USGS 1981). Note that the greatest increases for both the A2 and B1 scenarios are for the middle third of the century, perhaps because precipitation (for the A2 case) declines in the latter part of the century, and a reduced snowpack reduces the likelihood of a large contribution of snowmelt during a rainstorm. For the B1 scenario, the 100-year expected flood for the middle third of this century is projected to be 2.5 times the currently accepted value. These changes in flood frequency have major implications for stream morphology and habitat in the Tahoe basin, as well as for infrastructure, such as roads and bridges, and they support the assertion of Milly *et al*. (2008) that "stationarity is dead."

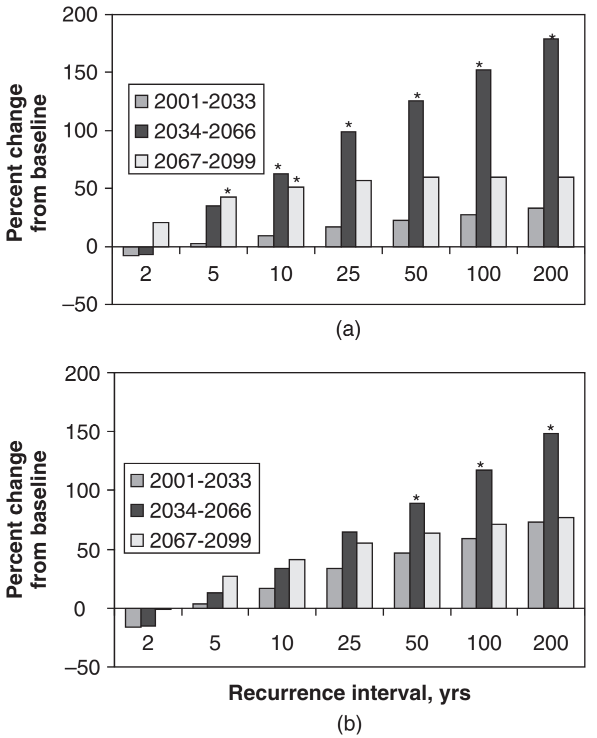


Figure 14.10 Percentage change in the modeled and adjusted GFDL/LSPC B1 and A2 flood frequency curves from the gauge record (1972–2008) (a) GFDL B1 and (b) GFDL A2.

14.4.4 Lotic ecology

Changes in stream flow regimes, especially an earlier and shorter snowmelt recession period, may have major effects on stream ecology. Possible effects (Yarnell, Viers and Mount, 2010) include: (i) a longer duration of the warm low-flow season; (ii) reduced suitable habitat and recruitment success for riparian woody vegetation; (iii) reduced arthropod diversity and changes in the macroinvertebrate community; (iv) flashier winter floods, with concomitant changes in sediment sorting and routing; and (v) greater habitat homogeneity and lower biodiversity.

14.5 Impact on the lake

Climate change and increased water temperature can have numerous effects on lake processes and biological communities. For example, Lake Tahoe has experienced a rapid increase in aquatic invasive species over the past decade, including bluegill sunfish (*Lepomis macrochirus*), largemouth bass (*Micropterus salmoides*), Eurasian water milfoil (*Myriophyllum spicatum*), curly leaf pondweed (*Potamogeton crispus*) and the Asian clam (*Corbicula fluminea*) (Kamerath, Chandra and Allen, 2008; Ngai 2008; Wittmann *et al*. 2010; TERC 2011; see also Chapter 15). These species inhabit the near shore region that is particularly vulnerable to temperature increase due to its shallow bathymetry.

The most important impacts of climate change on the lake, however, will be the direct result of lake warming on lake mixing, and in turn on internal nutrient loading and the lake's ability to support life in the hypolimnion. Figure 14.11 shows the modeled trends in volume-averaged, whole-lake temperature over the twenty-first century. For both scenarios, the warming trend was highly significant ($P < 10^{-6}$, by the Mann-Kendall test) but it did not really begin until well into the century (2039 for the A2 case and 2028 for the B1 case). In fact, the lake temperature under the A2 scenario dropped slightly ($P < 6 \times 10^{-4}$) from 2001 to 2039 before beginning a long upward trend.

The warming trend, especially under the A2 emissions scenario, will increase the lake's thermal stability and resistance to mixing, for two reasons. First, during summer, the lake warms from the surface downward, which increases the slope of the vertical temperature gradient; and second, the decrease in density of water with increasing temperature is highly non-linear. The work required to mix layered water masses at 24 and 25 °C is 30 times that required to mix the same masses at 4 and 5 °C (Wetzel 2001). The lake modeling results show that the lake's increased thermal stability will increase during the twenty-first century, suppressing complete lake turnover for prolonged periods of time (Sahoo *et al*. in press).

The Schmidt Stability (S) is a measure of the work required to mix a thermally stratified lake to an isothermal state without loss or gain of heat. It can be calculated from the lake's vertical density profile and bathymetry (Idso 1973). Figure 14.12 shows the lake's modeled minimum annual Schmidt Stability S (in $\mathrm{kJ\,m^{-2}}$). As with the temperature trends, the upward trends in annual minimum S for both scenarios were highly significant ($P < 0.002$), but were strongest over the last 60–70 years of the twenty-first century. Note in Figure 14.12 that the Schmidt Stability showed some persistence. If the value of S in a given year remains above $2.5\,\mathrm{kJ\,m^{2}}$, then the following year it never reaches 0. This persistence could be the result of the lake's thermal inertia, persistence in the modeled climate system, or both.

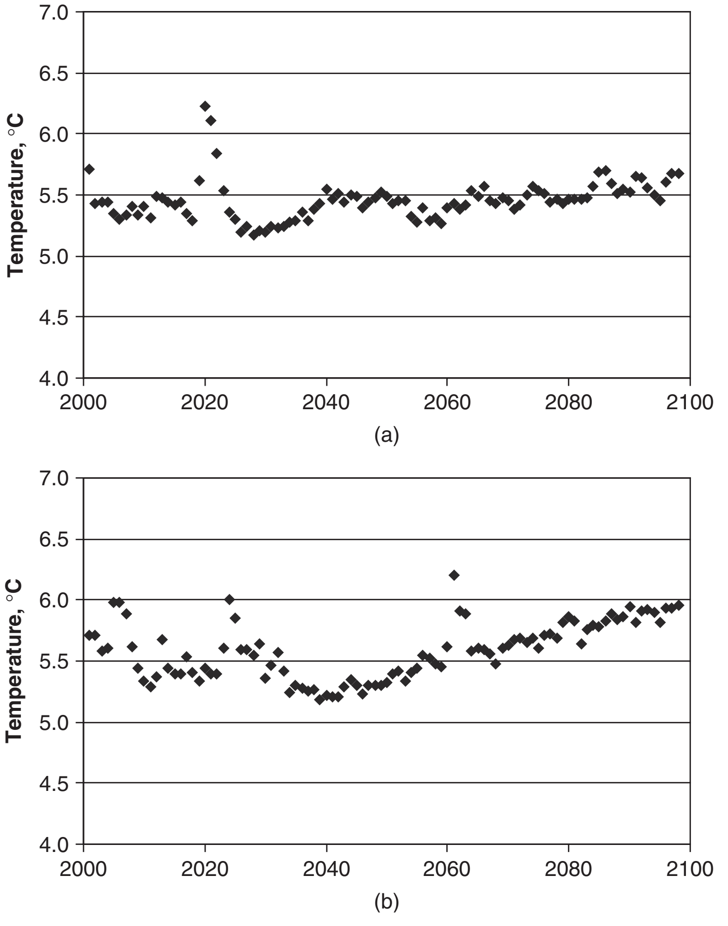
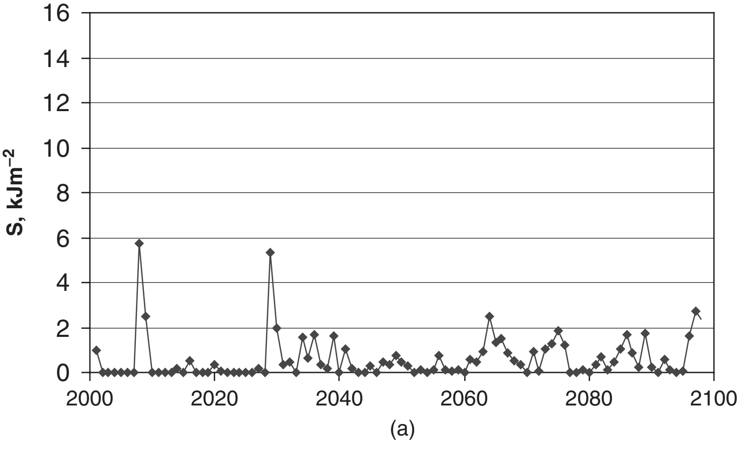


Figure 14.11 Modeled time trends of volume-averaged lake temperature (a) GFDL B1 and (b) GFDL A2.

Figure 14.13 shows the maximum mixing depth for the twenty-first century. Under the A2 scenario, the lake. mixed to the bottom only three times between 2069 and 2100. Under the B1, deep mixing occurred less frequently during the second half of the century, but did not cease altogether in this century.

The late spring mixing, however, was determined not just by the lake's thermal structure, but also by wind. The modeled GFDL A2 wind trends downward over the twenty-first century at an average of -0.067 +/− $0.045\,\text{cm sec}^{-1}\ \text{year}^{-1}$, according to the Mann–Kendall trend test. Sensitivity tests with the lake model showed that an increase of 10–15% in the modeled average daily wind speed over the twenty-first century would be sufficient to compensate for the effects of increasing temperature on thermal stability, in spite of the downward trend (Sahoo *et al*. in press). An evaluation of the modeled wind trends by month indicated increases in summer winds (which will tend to deepen the epilimnion) and decreases in the late fall and winter winds.

The current frequency of deep mixing—about one year in four—was sufficient to keep the lake well ventilated, and maintain oxidized conditions in the surficial



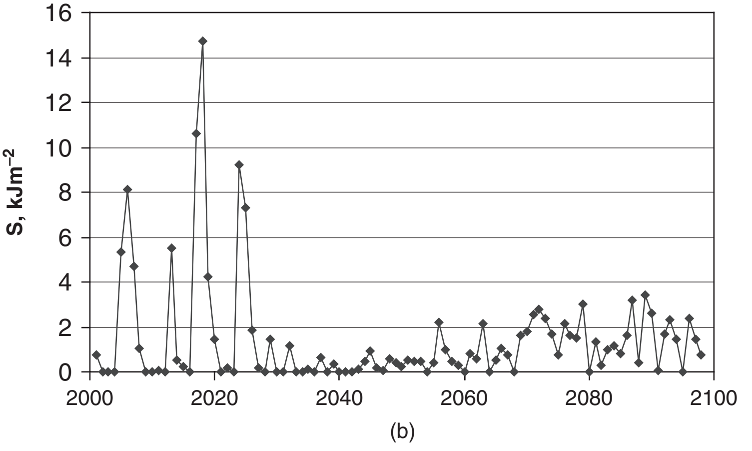


Figure 14.12 Modeled time trends in Schmidt Stability index (a) GFDL B1 and (b) GFDL A2.

bottom sediments. The loss of deep mixing would shut down the deep water ventilation, causing intense and prolonged periods of anoxia (Figure 14.14). This in turn will trigger large releases of soluble reactive phosphorus (SRP; Figure 14.15) and ammonium-nitrogen from the sediment into the bottom waters.

Beutel (2000) measured the release of available N and P from Lake Tahoe sediments, and his experiments provide a basis for calculating internal nutrient loading under anoxic conditions. This internal loading can then be compared to current estimates of external loading, which comprise chiefly watershed runoff and direct atmospheric deposition (LWRQCB and NDEP 2010). According to these calculations, by the end of the century under the A2 scenario, the internal load of SRP will be about twice the present external load, and the internal load of dissolved inorganic nitrogen (DIN) will be about one-third of the present external load.

The rate at which the released nutrients will be transported into the photic zone is currently unknown because active deep-mixing will cease as a result of the thermal stability in the water column. Initially the released nutrients will be trapped below the photic zone by density stratification, but eventually may be entrained into the upper

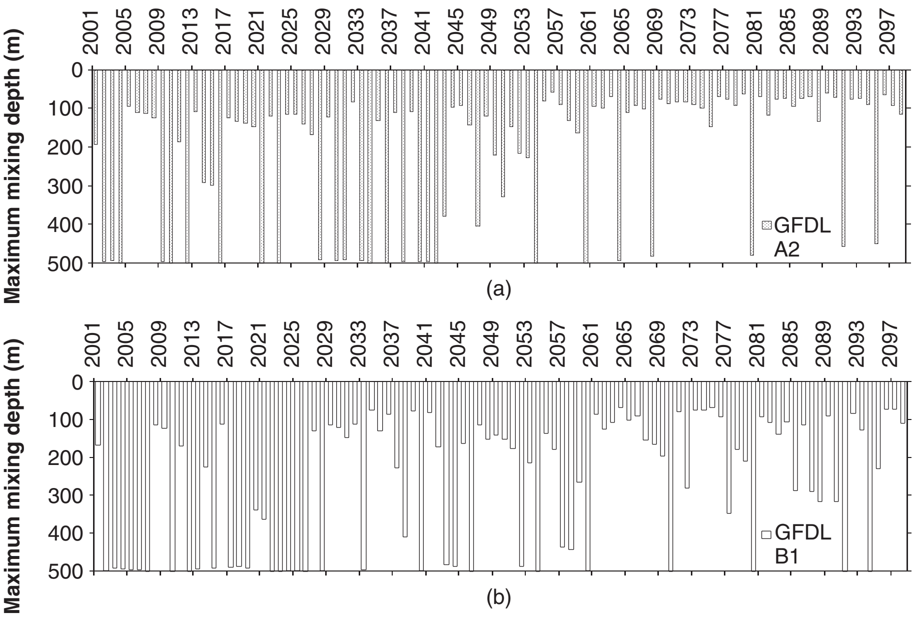


Figure 14.13 Maximum mixing depth of Lake Tahoe for two emission scenarios.

waters by vertical eddy diffusion. The net result would be a rapid increase in the lake's primary productivity and possibly a shift back to nitrogen-limitation because of the increased internal P-loading. These changes would most likely be irreversible, since the increased algal growth would increase the organic matter deposition in deep water, adding to the BOD of bottom waters and initiating, through a positive feedback mechanism, a "death-spiral of anoxia."

The development of anoxia and release of phosphorus from the lake sediments would trigger a cascade of microbial and biogeochemical changes (see also Chapter 6), though these are hard to predict. For example, denitrification in the sediments and water column could reduce available nitrogen, and hasten the shift back to nitrogen limitation. Blooms of cyanobacteria, however, could provide a new source of available nitrogen. While the pathways and rates of biogeochemical changes are uncertain, it is likely that Lake Tahoe will lose its unique ultra-oligotrophic status.

Release of methyl mercury from lake sediments and its incorporation into the food chain is another possible effect of anoxia (Watras 2010). In a study of heavy metal concentrations in the sediment of Lake Tahoe, Heyvaert *et al*. (2000) found that the Hg concentration in modern (post-1980) lake sediments was five times the pre-1850 concentration, and that the modern flux (mostly from direct atmospheric deposition) was 28 times the pre-industrial flux. Drevnick *et al*. (2009) found a lower modern to preindustrial flux ratio of 7.5 to ten, which was still relatively high compared with lakes world-wide. The concentrations of mercury in the tissue of lake trout (*Salvelinus namaycush*) and crayfish (*Pacifastacus leniusculus*) from Lake Tahoe showed a positive relationship between mercury concentration and weight, suggesting bioaccumulation. Consumption of one serving from the largest (10 kg) fish sampled (with a total Hg muscle tissue concentration of 0.24 $\mu g\,kg^{-1}$) could easily exceed EPA's recommended

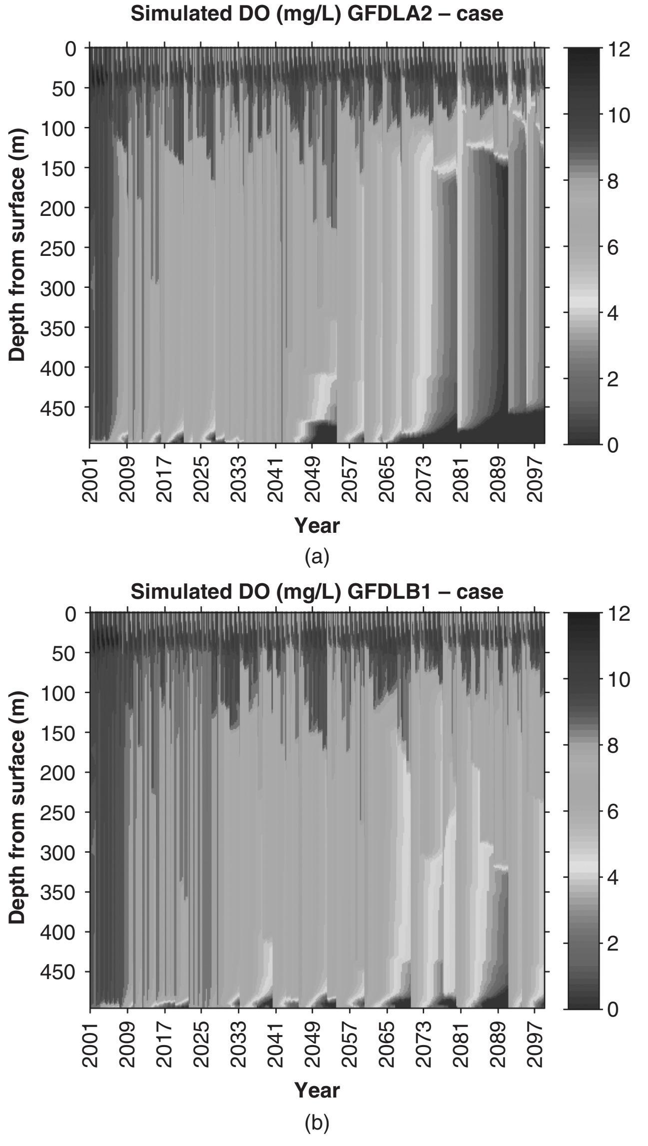


Figure 14.14 Twenty-first century dissolved oxygen trends in Lake Tahoe for two emission scenarios. (See insert for color representation.)

upper limit for daily intake of methylmercury (http://www.epa.gov/iris/subst/0073.htm, accessed July 11, 2012).

The change in the lake's thermal structure will affect the input of fine sediment and nutrients directly to the photic zone by modifying the "insertion depth" of streams flowing into the lake. This depth is determined largely by the relative densities of the inflowing streams and the receiving waters. Under current conditions, most of the inflow from January to March plunges into deep water below the photic zone, and has little immediate impact on primary productivity and Secchi depth. Future stream

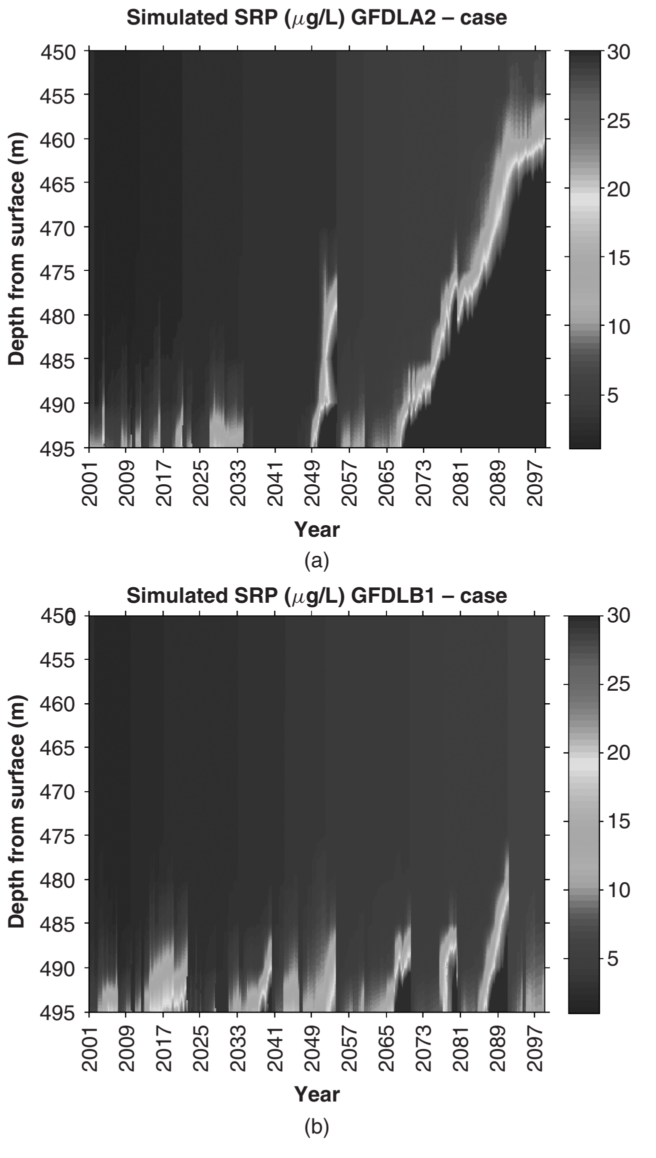


Figure 14.15 Twenty-first century internal loading of soluble reactive phosphorus in Lake Tahoe for two emission scenarios. The modeled pattern on NH_4-N release is the same as that of SRP, but the scale runs from 0–50 μg L^{-1} instead of 0–30 μg L^{-1}.

temperatures were modeled from air temperature and solar radiation. Comparison of the modeled stream temperature with lake temperature indicated that with the A2 scenario, the insertion depth of the Upper Truckee river will remain below 200 m, and out of the photic zone during the latter quarter of this century, whereas under the B1 scenario, the river discharge will more frequently deliver its sediment and nutrient loads directly to the photic zone (Sahoo *et al*. in press).

The suppression of deep mixing and consequent anoxia will have profound biological effects, making the deep waters of the lake uninhabitable for salmonids and many invertebrates. More subtle biological changes, however, are already under way at the lowest and highest trophic levels. Winder, Reuter and Schladow (2008) showed how warming of the surface waters and increased stability is shifting the phytoplankton population toward small-bodied more buoyant species and away from larger bodied species that sink more readily out of the photic zone and are more dependent on vertical mixing. As the lake warms and its clarity declines, it is becoming increasingly hospitable for the warm-water fish mentioned above.

Lake Tahoe is not only a priceless scenic, recreational and biological resource. In most years, it drains into the Truckee River, supplying water to municipal and agricultural users downstream in the Reno, NV vicinity and helping to maintain Pyramid Lake at the river's terminus. From 1910 to 2010, the Tahoe basin contributed on average 23% (ranging from 0–40%) of the annual yield of the Truckee River at Farad, near the CA-NV border. In 20 out of the last 110 years, however, the lake level has fallen below the natural rim. Under both the A2 and B1 scenarios, the lake level will fall below the rim sporadically for much of the twenty-first century, but recover during years of heavy precipitation. Under the A2 scenario (Figure 14.16), the lake level will drop below the rim in about 2085, and plunge to an elevation of 1894.1 m by the end of the century, a level 1.5 m below the historic (since 1900) low stand of November 1992. Once the lake drops below the rim, evaporation from the lake is cumulative from one year to the next. Combined with earlier snowmelt and the shift from snow to rainfall, the declining lake level will present a serious challenge to water resource managers in the Truckee-Carson basin.

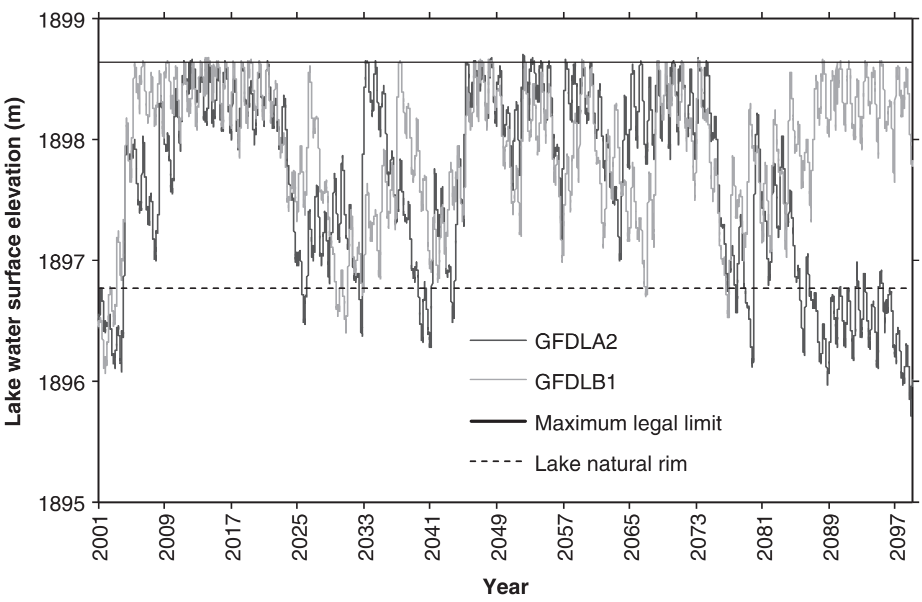


Figure 14.16 Modeled elevation of Lake Tahoe for two emission scenarios. When the lake level falls below the rim elevation, outflow ceases. (See insert for color representation.)

14.6 Reducing negative impacts

Climate change over this century threatens to shut off the deep mixing of Lake Tahoe, inducing anoxia and triggering a large release of algae-stimulating nitrogen and phosphorus from the bottom sediments. Changes in the watershed (described above) may increase the input of nutrients and fine sediment, reducing water transparency and promoting the reproductive success of invasive warm-water fish. At the regional and local levels, there is no way to halt the warming of the lake and to prevent the increase of internal nutrient loading—that solution must be found at the national and international levels. But it is possible to reduce the external nutrient loading to at least partially compensate for the future increase in internal loading, and watershed changes associated with climate change.

In 2010, the Lahontan Regional Water Quality Control Board and the Nevada Division of Environmental Protection completed their work on the "Lake Tahoe Total Maximum Daily Load Report," or TMDL (LRWQCB and NDEP 2010). Using a data base of 160 station years of stream gauging and sampling records, intensive sampling (450 events) of urban runoff, atmospheric deposition data, modeling of groundwater nutrient load from well data, stream channel erosion data, together with LSPC (Riverson 2010; Riverson *et al*. in press), the TMDL project calculated total loads of nitrogen, phosphorus and fine sediment from six source categories: non-urban uplands, urban uplands, atmospheric deposition (wet and dry), stream channel erosion, groundwater and shoreline erosion. Multiple regression was used to attribute nutrient and sediment loads to land use/land cover class (Coats *et al*. 2008).

Table 14.2 shows the final estimates of nutrients and sediment loads by source area. Note that direct atmospheric deposition is the most important source of total nitrogen, and upland urban areas are the most important sources of total phosphorus and fine sediment. The agencies have established a water transparency standard of 24 m Secchi depth after 20 years, and 29.7 m (the average transparency from 1967 to 1971) after 65 years. Using the Lake Clarity Model (Sahoo, Schladow and Reuter 2010), they found that the transparency standard could be met after 65 years by reducing the total loads of fine sediment, phosphorus and nitrogen by 65, 35 and 10%, respectively, from the currently estimated annual loads. If we assume that all of the current external load of phosphorus to the lake will ultimately be stored in the sediment and become available for release under anoxic conditions, then achieving the TMDL goal of a 35% reduction

Table 14.2 Pollutant loading estimates by source category, from Lahontan and NDEP (2010).

Source category		Total N, metric tonnes year^{-1}	Total P, metric tonnes year^{-1}	No. of fine sediment particles ($<16\,\mu$m), 10^{18}
Upland	Urban	63	18	348
	Non-urban	62	12	41
Atmosph. deposition (wet+dry)		218	7	75
Stream channel erosion		2	<1	17
Groundwater		50	7	NA
Shoreline erosion		2	2	1
Total		397	46	481

in load over the next 65 years could significantly reduce the internal load when anoxic conditions develop.

To achieve these goals, the TMDL project developed a broad array of Best Management Practices (BMPs) that could be used to reduce pollutant loads. Included are measures such as stabilizing and revegetating road shoulders, installing and maintaining storm water treatment vaults, detention basins, and treatment wetlands, putting to bed poorly-maintained dirt roads, and so forth. As with programs to reduce carbon emissions, no single measure can contribute a large load reduction, but each contributes a thin wedge, and taken together, the wedges add up to meet the program goals. The total estimated cost of construction and maintenance needed to achieve the 15-year transparency goal was estimated to be $100 million per year (LRWQCB and NDEP 2010).

The effectiveness (or design requirements) of some BMPs, however, may be affected by climate change, especially increases in rainfall intensity and rain-on-snow events. Wolfe (2010) used a combined water quality-runoff model—the Pollutant Load Reduction Model—together with the downscaled and bias-corrected twenty-first century precipitation (from GFDL with A2 and B1 scenarios) to investigate this problem. They found that increases in storm water runoff were reflected in a 10% decline in the treatment performance of facilities (such as treatment vaults and detention basins) designed for current hydrologic conditions. However, the designs based on current criteria would still be able to detain about 80% of the average annual runoff from developed areas. In general, the water quality benefits of implementing the BMPs in the proposed TMDL program would far outweigh the slight loss in functionality associated with climate change.

14.7 Summary and conclusions

Analysis of historic climate data in the Tahoe basin dating back from the early twentieth century indicated strong upward trends in temperature, especially for minimum daily temperature during spring and summer months. The upward trend was associated with a shift from snow to rainfall, a decrease in the number of days with average temperature below freezing, and a shift in the timing of the spring snowmelt peak runoff to earlier dates, and (since 1970) a warming trend in the lake itself. Comparisons of the rates of warming and the shift in snowmelt timing with stations nearby but outside of the basin suggested that the Tahoe basin is warming faster than the surrounding region.

The downscaled GCM output for both the B1 and A2 emissions scenarios indicated that the warming of the basin will continue. When we used the bias-corrected output for the GFDL model to force a hydrologic model over the twenty-first century, the results indicated a loss of snowpack, shift of the annual runoff hydrograph toward earlier dates, a large increase in flood risk in the middle third of the twenty-first century, and increased drought severity, due in part to the declining contribution of snowmelt to soil moisture storage. Changes in precipitation were less certain, though with the higher emission scenario (the A2), there was indication of decreasing precipitation in the last quarter of the century.

To evaluate the impacts of climate change on the lake, we used the output of the hydrology model together with downscaled climate data to drive a combined hydrodynamic/water quality model for the lake. The results indicated that, with the higher emission scenario (A2), the lake will continue to warm, especially in the latter two-thirds of the century, and become more thermally stable. Deep mixing will cease for long periods and the deep water of the lake will become anoxic, triggering a large release of available phosphorus and nitrogen, and a significant loss of well-oxygenated lake volume that now serves as salmonid habitat. By the end of the century, the lake may drop below the rim to an elevation not seen in historic time. Although these changes cannot be prevented without a large global reduction in future greenhouse gas emissions, the strict application of water quality Best Management Practices, though expensive, offers an opportunity to mitigate and perhaps forestall the potentially disastrous impacts of climate change on Lake Tahoe.

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