

Modelling surface–groundwater interaction in the Ruataniwha basin, Hawke’s Bay, New Zealand

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Abstract Modelling groundwater and surface water is important for integrated water resources management, especially when interaction between the river and the aquifer is high. A transient groundwater and surface water flow model was built for Ruataniwha basin, New Zealand. The model covers a long-time period; starting in 1990, when water resources development in the area started, to present date. For a better resolution, the simulation period was divided into 59 stress periods, and each stress period was divided to 10 time steps. The model uses data obtained from surface water, and groundwater collected over the last 20 years. Rivers and streams were divided into 28 segments and flow and streambed data at the beginning and end of each segment was used. Parameter estimation and optimisation ‘PEST’ was used for automatic calibration of hydraulic conductivity, groundwater recharge and storativity; whereas riverbed conductance was manually calibrated. Model results show that the rivers gain from the aquifer considerably more than the river losses. The cumulative groundwater abstraction over the last 20 years is approximately 210 million m³. This amount is very low compared to other water budget components; however, the effect of groundwater abstraction on storage is significant. Based on the results of this study, it was found that the loss of storage over the last 20 years is more than 66 million m³. Results also reveal that the effect of groundwater abstraction on rivers and springs flow is significant. The

rivers gain from the groundwater system, and the springs flow have been decreasing.

Keywords Surface/groundwater interaction · Groundwater modelling · Ruataniwha basin · Modflow · Stream depletion

Introduction

Rivers, springs and aquifer constitute an integrated system that should be managed as one entity. Maintaining a certain flow in the rivers and spring-fed stream is vital, not only for sustainable management but also for a healthy ecosystem. Managing aquifers and rivers separately is not valid, especially in high-interconnected aquifer–river systems. In addition to stream depletion, wells may intercept lateral groundwater flow that would otherwise end up in the nearby stream. It is important to understand the river–aquifer interaction for integrated water resources management.

Many analytical models have been developed to quantify the effect of groundwater abstraction on a nearby stream (Theis 1935; Glover and Balmer 1954; Hantush 1965; Hunt 1999, 2008; Butler et al. 2001). The problem with analytical models is that they are too simple to represent a regional heterogeneous surface/groundwater system. For example, these models assume a straight infinite river, and a steady pumping rate in an isotropic homogeneous aquifer. All these assumptions do not exist in reality. In addition, rivers may lose water to an aquifer at one location and gain water at another. The analytical models cannot cope with such a case (Rushton 2007).

Limitations of the analytical solution can be overcome using numerical models. This has been explored by

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Spalding and Khaleel (1991), as they compared the results of both solutions of a stream depletion problem. Due to the assumptions involved in the development of analytical solutions, they tend to overestimate the stream depletion resulting from pumping (Sophocleous et al. 1995; Fox et al. 2002).

Numerical models can overcome the limitations of the analytical solutions. They can also handle more complicated systems that the analytical solutions cannot. The only drawback of numerical models is that they require more data and processing time.

The Ruataniwha basin in Hawke's Bay, New Zealand, has come under increasing pressure in the last decade pursuant to an increasing demand for water. Rivers and springs in the basin have shown a significant decrease in flow, and some river reaches dry up each year in peak summer demand. To understand the dynamic of the surface/groundwater system in the area, an integrated three-dimensional surface/groundwater model has been developed using MODFLOW 2000 (McDonald and Harbaugh 1988). The Visual Modflow software version 2010 (Schlumberger Water Services 2008) has been used for this purpose. The stream and the drain packages of MODFLOW were used for stream routing and spring simulation, respectively. The model covers a time period of 20 years (from 1990 to 2009), and includes 12 rivers and streams (Table 1).

The study area

The area of study is located in the Hawke's Bay Region, New Zealand (Fig. 1). The surface catchment of Ruataniwha is approximately 1,470 km², while the study area extends over the basin, which has an area of 806 km². This comprises the gravel sediment formation from the late

Table 1 Rivers in the Ruataniwha basin (from north to south)

River/stream	Symbol
Mangamate stream	MM
Mangamauku stream	MU
Mangaonuku stream	MN
Waipawa river	W
Kahahakuri stream	K
Tukituki river	T
Avoca	A
Tukipo river	TP
Mangatewai stream	MG
Makaretu stream	MK
Porangahau stream	P
Maharakeke stream	MH

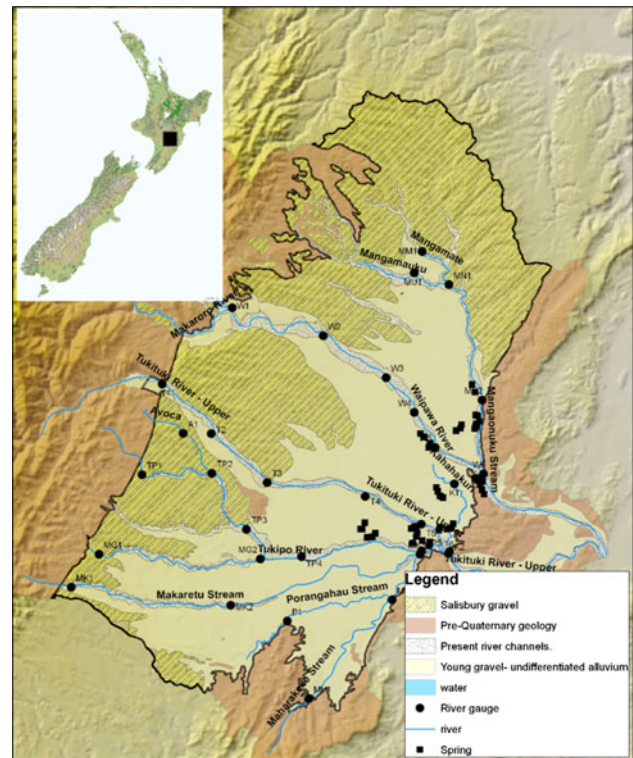


Fig. 1 The study area (top left), geology and model boundary

Pleistocene and the more consolidated gravel deposits (i.e. Salisbury Gravel) from the Pleistocene.

The basins' rivers and aquifers supply water for domestic, agricultural and industrial use, to the towns of Waipawa, Waipukurau as well as the smaller communities in the area. Despite the low domestic water demand in the area, due to the low population, the agricultural water demand (both current and projected) is high. The increasing demand on water for agricultural activities and dairy farming has resulted in stresses on water resources. Consent data from Hawke's Bay Regional Council suggests that the extent of irrigated land within Ruataniwha basin has increased from 260 to 2,200 hectares within the period between 1985 and 1995 (Dravid and Cameron 1997). The total area of irrigated land was estimated at 4,569 hectares in 2003 (Hawke's Bay Regional Council 2003), and is currently approximately 7,000 hectares.

Annually, the ecosystem of the basin is regularly placed at risk as many streams and springs dry up in the summer, resulting from peak demand for groundwater abstraction and long dry periods with little or no rainfall.

The basin geology is composed mainly of sequences of alluvial gravel from the Quaternary period with intermittent clay layers of different thicknesses (Fig. 1). Two main gravel layers occur in the basin. The top layer is comprised of young gravels from the late Pleistocene and Holocene epoch. It sits over an older gravel layer 'Salisbury Gravel'

from the early Pleistocene. The Young Gravel Formation is unconsolidated and contains clay, silt and volcanic ash of late Quaternary (Francis 2001). It is more permeable than the Salisbury Gravel underneath, which is comprised of slightly too poorly consolidated gravel, ignimbrite and clay from the lower Quaternary (Francis 2001). Intermittent volcanic activities in the central north island throughout these regions have resulted in the deposition of volcanic ash, pumice and ignimbrites over wide areas of the region.

Several rivers and streams traverse the basin from west to east and join just out of the basin to form the Tukituki river. The average annual inflow volume into the basin through the main three rivers; Waipawa, Tukituki and Makaretu is 550 million m³. The total average rivers' outflow is approximately 940 million m³/year.

With rainfall values of more than 1,000 mm/year, the basin receives a great portion of recharge from rainfall. In addition to rainfall recharge, there is a strong interaction between surface and groundwater. Historical flow records indicate that the rivers lose water to the groundwater in the western part of the basin and gain water from the groundwater in the eastern part of the basin.

The Ruataniwha basin groundwater system is being recharged by rainfall and surface flow (rivers). This recharge starts in the west of the basin at Ruahine ranges. The ranges receive a high rainfall (>1,300 mm/year) and because of their steep topography, most of this rain converts to surface runoff and enters the basin as rivers and streams. The average annual volume of rainfall in the basin is approximately 850 million m³.

Model development

Conceptual model

The hydrogeology of the area is complex and most of the aquifers are unknown. Based on the geological investigations (Pattle Delamore Partners 1999; Francis 2001), the subsurface geology was classified according to the age of formations. These formations are the Young Gravel Formation at the top and the older gravel formation (Salisbury Gravel Formation) underneath. Many aquifers occur in both formations. Based on this information, most shallow wells (<40 m deep) are unconfined, whereas all deep wells are confined. Therefore, the model was divided into three layers of variable transmissivity and storativity. Data from more than 50 aquifer test, obtained from the Hawke's Bay Regional Council database, has been used to assess hydraulic properties. In addition, data from more than 400 well logs and previous studies have been used to get a good estimation of

the hydraulic properties (Murray 2002a, b; Phreatos 2003). In the calibration process, hydraulic properties have been adjusted to get a good match between simulated and measured heads.

According to aquifer tests, the top layer is more permeable, less consolidated and mostly unconfined. The lower layer is less permeable, contains more clay and mostly confined. The basin is, however, very heterogeneous, and it is assumed that the aquifers are connected and that the entire basin is considered as a single entity in the context of water resources management. The thickness of each layer increases gradually from the edge of the basin towards the centre with both layers reaching a thickness of about 200 metres in the middle of the basin.

The Ruataniwha basin is a closed system; that is, there is no interconnection between the basin and other groundwater bodies. Water flows into the basin by means of surface flow via streams and rivers, in addition to recharge from rainfall. The model outer boundaries are considered as no-flow boundaries as the lateral groundwater flows into or out of the basin are thought to be either negligible or non-existent. All the cells within the model boundaries are active cells and were used during simulations.

Model discretization

The total area of the model is 806 km², which covers the gravel sediment area and the rolling hill country to the north of Mangaonuku river. A model grid of 97 rows, 83 columns and three layers was used to represent the different geological formations. Each grid cell has a dimension of 500 × 500 m. The total number of cells in each layer is 8,051, of which 3,211 cells are active. The total number of active cells in the entire model is 9,633.

The transient model covers a time period extending from 1990 to 2009. This simulation period was divided into several stress periods based on seasonal changes to cope with summer/winter variations, and also based on availability of data. The model synchronises the stress periods based on summer/winter abstraction variation and rivers flow variation.

Abstraction data was divided into three periods every calendar year, following summer and winter variation. These periods are from January to March, from April to September and from October to December. The total number of stress periods is 59, and each stress period is divided into 10 time steps of which the longest time step is 18 days.

Boundary conditions

The model area was considered as a closed basin, i.e. there is no lateral groundwater flow out of or into the

basin. The only means of groundwater flow is via the surface water outflow at the eastern boundary of the basin by the Waipawa and the Tukituki rivers. The outer boundaries of the basin were treated as no flow boundaries (Fig. 1).

Rainfall recharge

Groundwater recharge is mainly from rainfall, with average annual rainfall varying between 800 and more than 1,300 mm. Rainfall recharge was estimated based on a previous study that was based on a water balance model (Baalousha 2008). The groundwater recharge was computed annually based on water balance and was distributed spatially based on soil properties, topography and rainfall spatial distribution. The average annual groundwater recharge from rainfall is 250 million m³ (or 310 mm/year).

Stream boundaries

The main rivers within the Ruataniwha basin are Waipawa river, Tukituki river, Tukipo river and Makaretu stream. The smaller streams are Mangaonuku Stream, Mangaoho Stream, Mangatewai, Avoca, Kahahakuri, Porongahau and Maharakeke. All rivers and streams in the basin join in two rivers; the Waipawa and the Tukituki. These two rivers join into one river about 10 km to the east of the basin.

For modelling purposes, rivers and streams were divided into segments. Division of rivers and streams into segments was based on data availability and river flow pattern. Figure 1 shows the surface water gauging sites where segments connect. The 12 rivers included in this model were divided into 28 segments.

The flow data at the upstream points, namely, MM1, MU1, W1, T1, A1, TP1, MG1, MK1, P1 and MH1 were inputted into the model. The model then simulates the flow and river/aquifer interaction at points downstream along each river.

Springs flow

There are many springs and spring-fed streams in the basin. The flow of these springs varies between 1 and more than 500 l/s (based on historical spring survey data).

The majority of the springs occur in the lower part of the basin where the groundwater potential is high. The drain package in Modflow has been used to simulate springs flow. Data required for springs' simulation includes the head of the spring and the conductance. Heads of springs were obtained from field measurements, and conductance was approximated at 1,000 m²/d based on measured springs flow.

Hydraulic properties

The hydraulic properties of the basin are not well researched, and few reliable investigations have been conducted in the area. Many aquifer tests; however, have been carried out in the basin over the last 30 years. Aquifer tests data have been used to assign the permeability at different locations in the basin.

The three layers representing the three aquifer formations mentioned above have been divided into different zones. The zonation was based on the surface subsurface geology, lithology and aquifer tests' data. Long groundwater time series allow for a better calibration of hydraulic properties. Figures 2, 3 and 4 (ESM only) show zonation of layer 1 (the shallow layer), layer 2 and 3 (the deep layer), respectively. Altogether, the three layers were divided into 12 zones. The calibrated hydraulic conductivities of these zones are shown in Table 2.

The basin was assumed to be isotropic and homogenous with a vertical anisotropy of 0.1, i.e. the vertical hydraulic conductivity was assumed to be 0.1 of the horizontal hydraulic conductivity.

Results of aquifer tests suggest that the area around the Waipawa river has the highest transmissivity, whereas areas in the southern part of the basin have the lowest transmissivity. Tritium data confirmed this assumption (Pattle Delamore Partners 1999).

The hydraulic conductivity values were selected based on previous modelling results (Baalousha 2009) and were adjusted in the calibration process.

Specific yield (for unconfined aquifers) and specific storage (for confined aquifers) are required for the transient simulation. Values of storage parameter can be obtained from the aquifer test analysis. Initial values of storage parameters were assigned based on aquifer tests results in the Ruataniwha basin and theoretical values from textbooks (Freeze and Cherry 1979; Fitts 2002). According to these books, storage coefficient values vary between 0.005 and 0.00005. The specific yield values vary between 0.1 and 0.3. The results of the steady-state model (Baalousha

Table 2 Calibrated values of hydraulic conductivity at different zones

Zone	Calibrated <i>K</i> value (m ² /d)	Zone	Calibrated <i>K</i> value (m ² /d)
1	19	7	0.7
2	3.3	8	1
3	60	9	60
4	51	10	35
5	1	11	60
6	4.1	12	42

2009) have been used as initial head for the transient model.

Groundwater abstraction

Estimation of groundwater abstraction was based on the crop water demand estimation and interpolation of the compliance data, which covers part of the pumping wells. Prior to 1995, there has been no data on groundwater abstraction as the pressure on water resources in the Ruataniwha basin was minor.

The volume of annual groundwater abstraction varies between approximately 3 million m³ in 1990 to 24 million m³ in 2009 (Fig. 5, ESM only). The cumulative groundwater abstraction is 210 million m³.

Model calibration

Model calibration is a process in which uncertain model parameters are changed to fine tune the model output and to achieve the best match between modelled results and field measurements of both groundwater levels and stream flow. The model parameters, which can be altered to produce a calibrated model, are the hydraulic properties and groundwater recharge.

Hydraulic properties include hydraulic conductivity, storativity and streambed conductance. While hydraulic properties can be considered constant over the entire simulation time, recharge varies from one time step to another.

Data from the state of environment ‘SOE’ monitoring wells were used to calibrate the model. The ‘SOE’ wells have time series of the groundwater head of variable periods over the simulation time period. Some wells have data from early 1992 up to date, while other wells have shorter time series. In addition to the SOE wells, data from a piezometric survey, undertaken in 1997 (Dravid and Cameron 1997), were used to cover any shortage in spatial wells distribution. More than 50 wells have been used for calibration.

Groundwater recharge, hydraulic conductivity and storativity were calibrated automatically using parameter estimation program (PEST) (Doherty 2009). No prior information was used in PEST. PEST settings include Marquart Lambda, which was given an initial value of 10, phi ratio of 0.3 and maximum trial lambdas of 10. For more information about these parameters, the reader is referred to Doherty (2009). Calibrated groundwater recharge values vary from 170 million m³/year (211 mm) to 360 million m³/year (447 mm). Calibrated hydraulic conductivity values vary from 0.5 to 60 m/d, as shown in Table 2. Streambed conductance was calibrated manually, and the modelled stream–aquifer interaction was examined against field measurements.

The goodness of calibration can be measured through the discrepancy in water budget, mean value of the residual errors and their probabilistic distribution. Discrepancy in water balance should be as small as possible. The maximum absolute discrepancy is 130 m³ or 0.004% of the water budget, which is negligible.

Figure 2 shows results of simulated compared to measured flow at the lower end of the Waipawa river (point W6) and the Tukituki river (point T6). As shown in the figure, 40 flow measurements over the simulation period have been used. The correlation coefficient (r^2) between measured and simulated flow at lower Waipawa and Tukituki is 0.987 and 0.982, respectively.

Figure 3 shows the simulated and observed head at some observation wells in the basin over the entire simulation period. It is clear there is a good fit between simulated and measured head over time. Although, the seasonal heads variation is not fully captured by the model because annual recharge was used and not seasonal; however, the mean head trend is well reproduced.

Results and discussion

The transient flow model for groundwater and surface water is an invaluable tool for water resources management as it answers many question regarding surface water and groundwater. The output of the transient model includes transient water budget, stream/aquifer gain and loss relationship and impact of pumping on water system dynamic. In addition to model output, sensitivity analysis shows which parameters are more important to model results than others, and thus, shows which parameters needs to be investigated further.

The histogram of residual error is shown in Fig. 8 (ESM only). The distribution of error is clearly following the normal distribution, with the mean value of -0.05 , which is close to 0. This is another measure of the goodness of calibration.

Groundwater level and flow direction

Groundwater head in the basin varies from 140 m at the eastern boundary to more than 300 m above mean sea level at the south–west boundary. The flow and velocity vectors are shown in Fig. 4. The flow is essentially from north–west to the south–east, with some local variations. Darcian groundwater flow velocity is expressed with the length of velocity vectors in the figure. The longer the vector is, the higher the velocity.

The gradient of groundwater (flow direction) is expressed by arrows (Fig. 4). Groundwater flow velocity is low upstream, and then increases as it travels across the basin,

Fig. 2 Simulated and measured flow at lower end of the Waipawa and Tukituki rivers

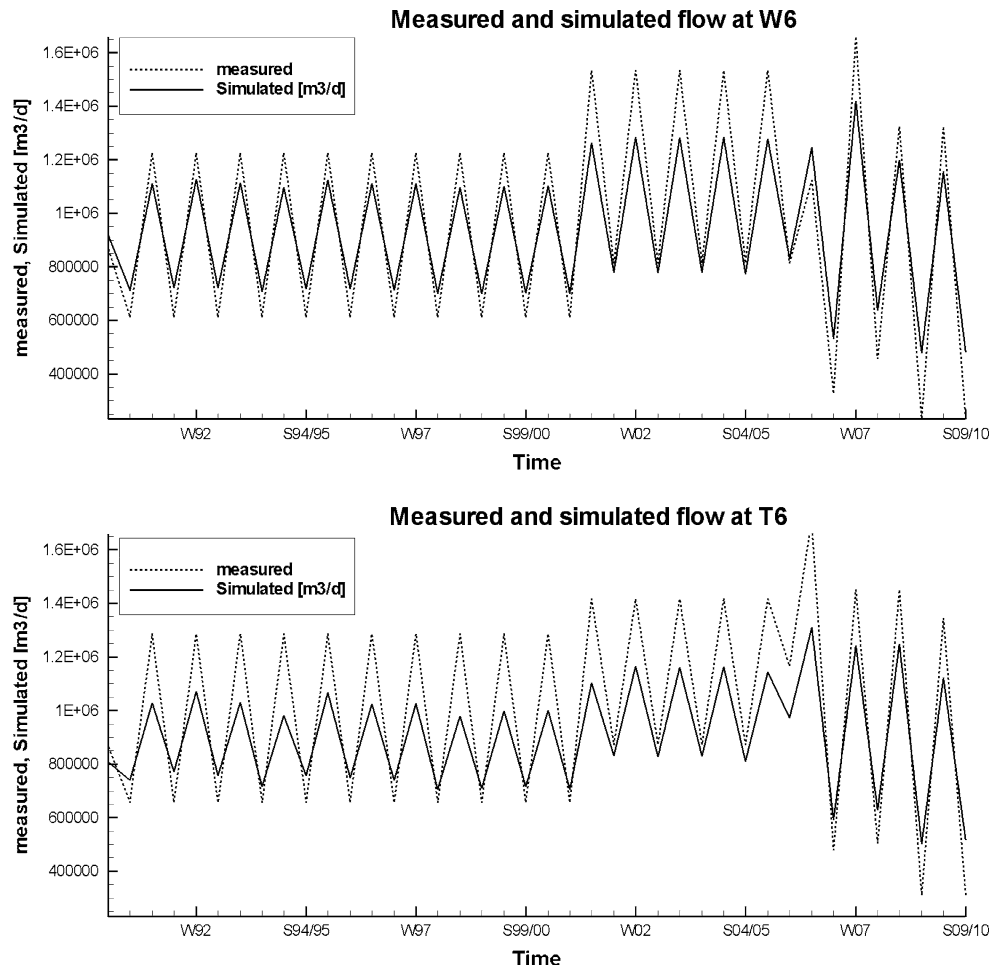
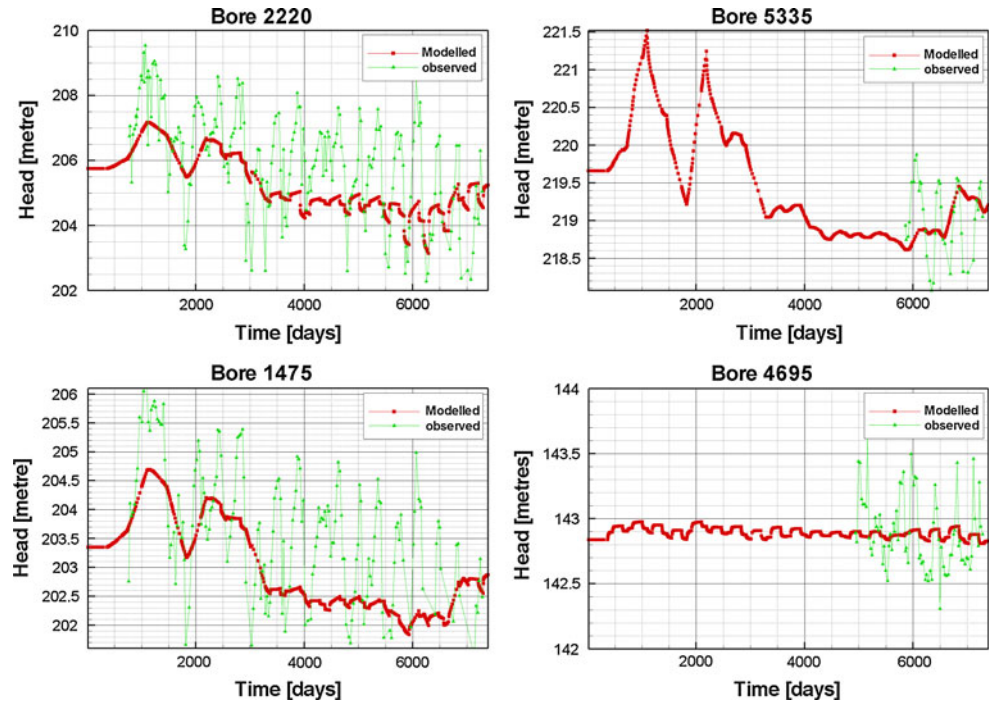


Fig. 3 Simulated and measured head at monitoring bores, over the simulated time period



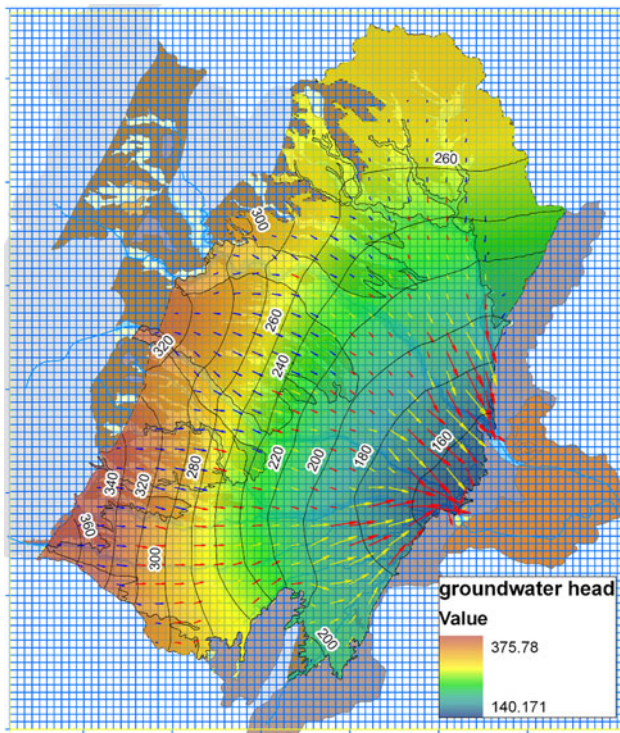


Fig. 4 Flow direction and groundwater head at the end of simulation period

and reaches its maximum speed at the eastern part of the basin. The relatively high speed there is because the transmissivity is high in the lower part of the basin, compared to elsewhere. At the eastern edge of the basin, and because the basin is closed i.e. there is no lateral groundwater outflow, groundwater has only two exit points, through the Waipawa and the Tukituki rivers.

It is also clear that the groundwater movement upstream in the hill country and rolling hill between Waipawa and Tukituki is low. This is because of the low permeability of the limestone hills in those areas.

Drawdown

Drawdown at the end of the simulation period, with respect to initial water level at the beginning of simulation, is shown in Fig. 5. The highest drawdown occurs around the Takapau area in the south, Ongaonga and Tikokino in the north, and upstream on the Mangaonuko river. Heavy groundwater abstraction in the last few years has taken place in these areas, which could explain why there is a high drawdown there.

Takapau and Tikokino are located at the southern and northern edges of the basin, respectively. In addition to high pumping at these areas, the groundwater flow at both areas has a down gradient. This downward movement also contributes to the high drawdown.

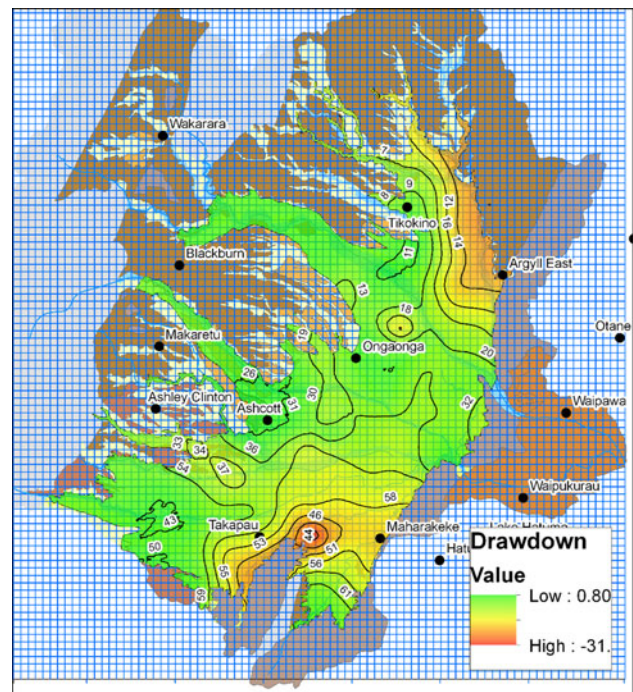


Fig. 5 Drawdown in metres at the end of simulation period

Water budget

The system is very dynamic as most components change with time. Clearly, the groundwater recharge from rainfall is highly correlated to stream gain from the groundwater system. The stream gain and rainfall recharge are the largest components in the water budget.

The overall water budget at the end of the simulation period is shown in Table 3. The main inflow component into the basin is the rainfall recharge, which constitutes 85.5% of the overall budget. The second largest component of inflow is the stream leakage (i.e. losses from streams to groundwater system), which is approximately 8.4% of the total budget. Storage losses constitute approximately 6.08% of the budget.

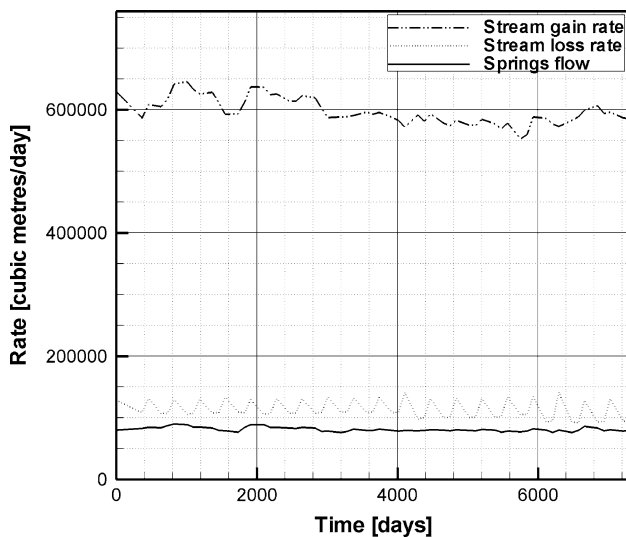
Flow components out of the basin are listed in Table 3. The largest outflow component is the groundwater losses to rivers, which are 81.07% of the total budget. Other components include the flow to drain (springs) and to constant head (in the steady-state simulation period). The loss to storage (or the storage gain) is approximately 4.99%. Wells abstraction constitutes only 3.53% of the total budget.

Groundwater–surface water interaction

The gain/loss relationship between aquifers and rivers is complicated and varies spatially and temporally. In some locations rivers gain from aquifers, in others they lose, but the overall budget of the entire basin shows that the total

Table 3 Cumulative water balance volume and percentage at the end of simulation period

Component	In (volume m ³)	Out (volume m ³)	In (% total)	Out (% total)
Storage	368,113,984	302,097,792	6.081	4.99
Wells	0	213,938,128	0	3.53
Drains	0	629,732,992	0	10.40
Recharge	5,174,923,776	0	85.50	0
Stream leakage	509,574,784	4,906,824,192	8.42	81.07
Total	6,052,612,608	6,052,593,152	100	100
In–out	19,456		0.00032	0.00032

**Fig. 6** Rates of streams and springs loss/gain

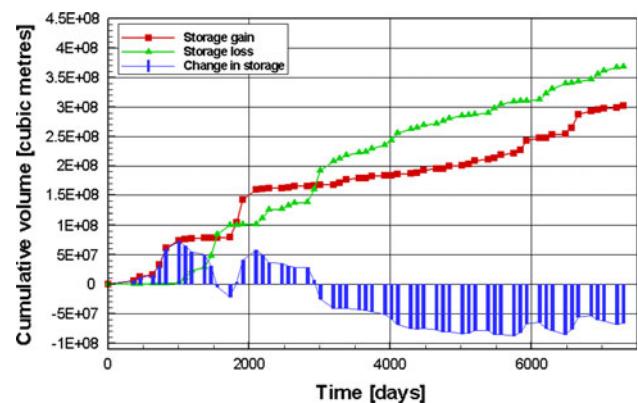
river gain from the groundwater is much larger than the loss. Figure 6 shows the rate of streams and springs loss and gain. The amount of river gains calculated over the entire simulation period of 20 years is 4,059 million m³, while the total loss is only 809 million m³ over the same period. Based on these results, it can be concluded that rivers gain most of their water as they move downstream.

It is important to note that there is a clear decreasing trend in river gain, as shown in Fig. 6. In the first 3,000 days (around 8 years), the average rate of river gain is approximately 613,544 m³/day. This rate decreased to 561,738 m³/day in the following years, which may be the result of increased pumping rate in the last 10 years. In other word, the decrease in rivers flow resulting from groundwater abstraction is approximately 650 l/s.

On the other hand, stream losses to groundwater system, although small, have been slightly increasing over the last decade, as a result of groundwater abstraction.

Storage losses

Groundwater abstraction in the Ruataniwha basin is low, compared to inflow and outflow components such as

**Fig. 7** Change in storage since 1990 till present

rainfall recharge or river flow. Despite this, abstraction has a considerable impact on the groundwater storage. It is well known that once groundwater development starts, the natural balance of the system is disturbed. This causes the storage to change until a new equilibrium is reached.

Figure 7 shows the cumulative change in groundwater storage over the past 20 years from 1990 until present. Storage loss means the system takes water from storage, and storage gain means the system gives water to storage or the storage increases. Obviously, there is a cumulative discrepancy or the difference between storage gain and loss. This storage loss is approximately 66 million m³.

Given the potential increase in water demand, it is likely that the loss in storage will continue until the water resources are fully utilised and until enough time elapsed for the system to reach a new equilibrium.

Spring flow

Water resources development in the basin had a clear impact on spring's flow, as shown in Fig. 6. Springs flow vary over summer and winter and from one stress period to another. There is a clear decline in spring's flow over time pursuant to groundwater development. Decline in flow rate took place after 3,000 days (approximately 8 years). In the period from 1990 to 1998, the average spring's flow rate was 84,000 m³/day. Afterwards, the springs' flow has

dropped to 80,000 m³/day. In 2008, the groundwater recharge was high due to above-average rainfall, resulting in an increase in springs' flow, as appears in Fig. 6.

Conclusion

A transient three-dimensional groundwater/surface water flow model has been developed for the Ruataniwha basin. The model uses the steady-state model results (Baalousha 2009) as initial groundwater conditions for the transient simulations. The main objectives of the transient model were to understand the groundwater/surface water interconnection and the dynamics of water resources, specifically to understand the effect of groundwater abstraction on rivers and stream flow, and to run different management scenarios.

It is believed that steady-state conditions existed prior to 1990, as little groundwater abstraction took place at that time. Groundwater pumping has gradually increased from 1990 to 2000, but the volume pumped has doubled in the last decade. The current annual groundwater abstraction is estimated at 24 million m³. The model spans a time period from 1990 until present. The model considers summer and winter variations of rivers flow and groundwater abstraction, to get a better resolution of model output.

Values of hydraulic properties were based on aquifer tests results and were refined during calibration. The model was calibrated using the monitoring results of water level from the state of environment 'SOE' monitoring network, in addition to historical piezometric survey results.

The calibrated model shows a good match between the measured and simulated head at the SOE wells. The model shows also good match between rivers measured and simulated flow.

Results of the transient state model reveal that the groundwater flow direction is from north–west to south–east. Groundwater flow gradient is downward upstream and upward downstream, as the basin is closed and the only means of outflow is the rivers. Groundwater head varies between 140 m (above mean sea level) downstream to over 300 m upstream. The highest drawdown, as a result of pumping, occurs at Takapau, Tikokino and Ongaonga townships.

Rivers and streams are in general gaining water from the groundwater system, and the amount of gained water increases as rivers pass across the basin. River's losses to the groundwater system are small compared to river gain. Water budget results show that the average groundwater recharge is approximately 256 million m³, which is almost equal to rivers gain from aquifers.

The transient water budget results show that river gain from the groundwater system has been decreasing over the

last 5 years because of increasing groundwater abstraction. Groundwater abstraction has also affected the groundwater storage resulting in a loss of 66 million m³ over the simulation period.

Sensitivity analysis revealed that the model results are highly sensitive to hydraulic conductivity, especially in the middle of the basin around lower Waipawa and less sensitive to storativity. Streambed conductance is important in terms of water budget, but the model is not sensitive to changes in conductance values.

Given the high river/groundwater interaction, it is recommended to manage both groundwater and surface water together within the framework of integrated water resources management. As rivers gain water from the groundwater system, pumping groundwater affects the rivers gain. It is important to carefully select locations of any pumping well, as the spatial distribution of pumping affects the river loss/gain pattern.

There is a need for more investigation on hydraulic conductivity, as the model results are more sensitive to changes in hydraulic conductivity than any other parameter. More investigation on springs and springs-fed streams is necessary as little data is available on springs.

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