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# River sediment salt-load detection using a water-borne transient electromagnetic system

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## Abstract

The salinisation of major river systems in Australia, and in other countries, is primarily determined by the upward movement of saline water from regional aquifers into the river. The migration has been accelerated due to irrigation schemes and farming practices that have changed regional hydraulic gradients driving saline-water in aquifers towards the major drainage points in the landscape. In this paper, we describe results from a transient electromagnetic (TEM) system that has been deployed to monitor the influx of saline water through sub-riverbed sediments. The deployment was a floating arrangement of a commercial fast sampling (high resolution) TEM system that is sensitive to shallow (<50 m depth) resistivity variations.

The technique has been extensively trialed around the River Murray town of Waikerie in South Australia. An initial series of surveys along a 40 km section of river showed a range of sub-riverbed resistivities between 1 and 20  $\Omega\text{m}$ , with a top layer of about 10–15  $\Omega\text{m}$  closely following the water depth. Regions of high-resistivity in the riverbed sediments correlated well with saline-aquifer borehole pumping locations, indicating a localized drawdown of fresher river water. Low-resistivity anomalies have been interpreted as regions of saline water influx into the river. The technique is now used for routine mapping in Australia, with over 800 km of the Murray surveyed, and has potential application to other major world river systems.

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## 1. Introduction

Salinity is an expensive nationwide problem in Australia. More than \$130 million of agricultural production is lost annually due to salinity, more than \$6 million is spent each year on building maintenance related to salinity in South Australia and \$9 million

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damage is caused to roads and highways in New South Wales alone (National Action Plan for Salinity and Water Quality, 2003; <http://www.napswq.gov.au/>). Estimates show that within 20 years, the city of Adelaide's drinking water will fail World Health Organisation standards for desirable drinking water on two out of every five days, due to the increasing salinisation of the River Murray (National Action Plan for Salinity and Water Quality, 2003).

Salinity problems in Australia are generally produced by an imbalance between the volume of water reaching the ground (by rainfall and irrigation) and that leaving the ground (by such processes as evaporation, transpiration or plant use, aquifer recharge and surface run-off) (Jones et al., 1994). The major causes of such an imbalance are excess irrigation and reduced transpiration (due to the clearing of deep-rooted plants). An excess of groundwater builds up, dissolving and mobilising sediment-stored salts. Once mobilised, these salts may generate salinity problems as the groundwater rises to the surface, or migrates laterally.

Salt-load monitoring techniques are required along the River Murray to determine locations where saline-water accession is highest. Such information is routinely used to design salt-interception schemes and to assess the efficiency of existing mitigation schemes. Monitoring techniques need to locate salt-load "hot-spots" on a scale of 50m, and must be fast and cost-efficient so that the survey can be repeated regularly.

In this study we have adapted an existing Transient Electromagnetic (TEM) system for deployment over water in order to study the specific hydrogeological problem of saline water accession into the River Murray. Hydrogeological problems have employed TEM techniques in the past. Fitterman and Stewart (1986) developed some theoretical basis to support the use of TEM for groundwater investigations through numerical models. They suggest a number of groundwater problems to which TEM might be applied. Most of these problems can be investigated with ground based systems (e.g., Poulsen and Christensen, 1998; Christensen and Sørensen, 1998; McNeill, 1990), but some special problems require surveys over water bodies. The high conductivity of sea water generally rules out off shore surveys except in the case of grounded transmitters and/or receivers (Cairns et al., 1996) but floating systems can be deployed over

freshwater bodies such as lakes and rivers. Hurwitz et al. (1999) have described results obtained from a floating TEM system first described by Goldman et al. (1996). Their system investigated between depths of 10 and 100 m. In a study of aquifer recharge Butler et al. (2004) employed the Geonics EM31 and EM34 (Frequency domain EM systems) which allowed efficient mapping of horizontal variations in apparent conductivity, but this allowed no vertical resolution of conductivity. DC Resistivity arrays have previously been employed in floating configurations (Allen and Merrick, 2003; Dyck et al., 1983; Amimoto and Nelson, 1970). However, the length of the array and a requirement for the array to be straight introduces logistical difficulties. Furthermore, better vertical resolution makes TEM a more favourable solution.

Our system is used to investigate the conductivity of the top 5 m of alluvial sediments, which responds primarily to the salinity of the water that they contain. If fresh water is present in these sediments, it is reasonable to assume that the regional hydraulic head is low, and the river "loses" water to the sediments. On the other hand, where sediments contain water of salinity higher than that of the river, the hydraulic head is larger and driving water from aquifers upwards. It is important for the system to be practical, cost-effective and rapid. Existing techniques that involve river-water sampling (known as "run-of-river") require at least five days of repeated measurement to complete and lack the spatial resolution to delineate localised salt hot spots. A high saline-water influx can be localised to a scale of 50 m or less (due to faults, clay lensing or trench locations in the river), compared with ~1 km resolution from the run-of-river.

## 2. Hydrogeology and salinity in the Murray Basin

The Murray Basin in Fig. 1 is a shallow sedimentary basin covering approximately 300,000 km<sup>2</sup> of southeastern Australia (Brown, 1989; Brown and Stephenson, 1991; Lukasik and James, 1998). Low-permeability layers exist between the major geological groups that act as confining layers for groundwater systems. The hydrology of the Murray Basin is divided into three major aquifer systems that correspond to the division of major sedimentary groups. This project focuses on the hydrogeology associated with the Waik-

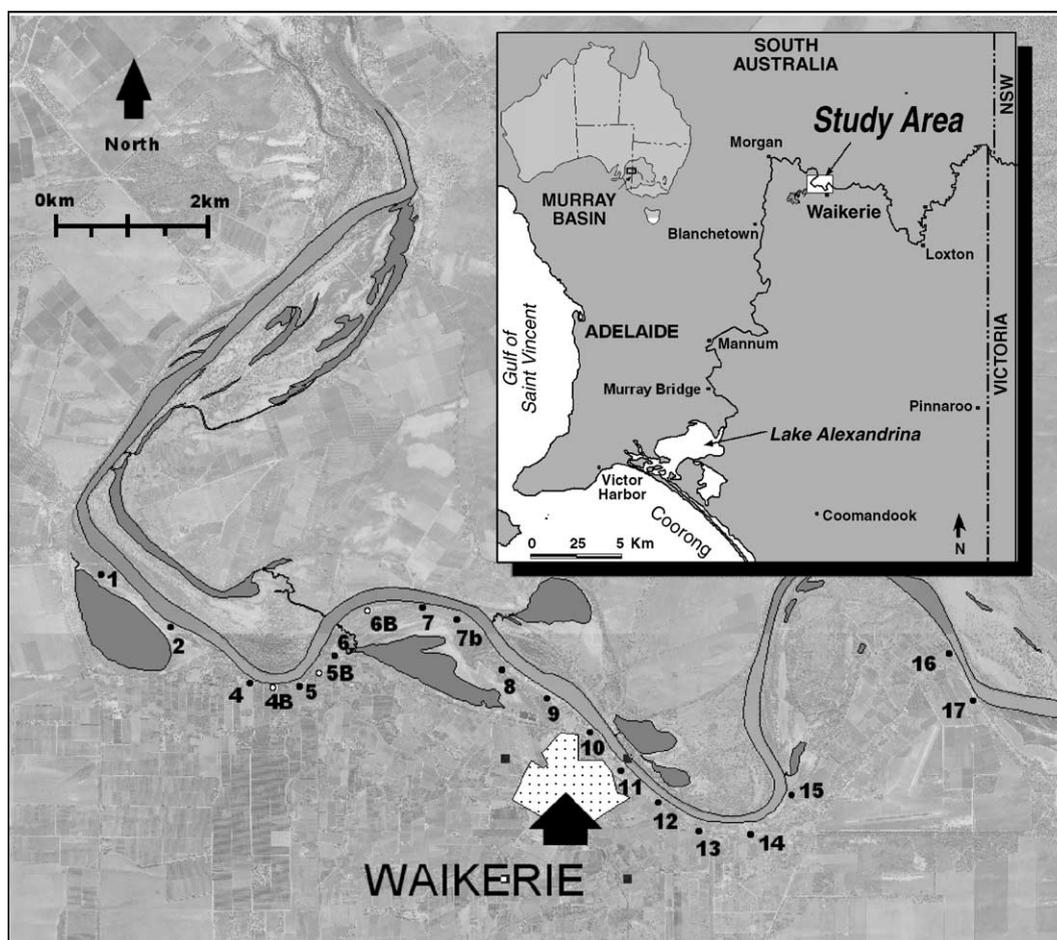


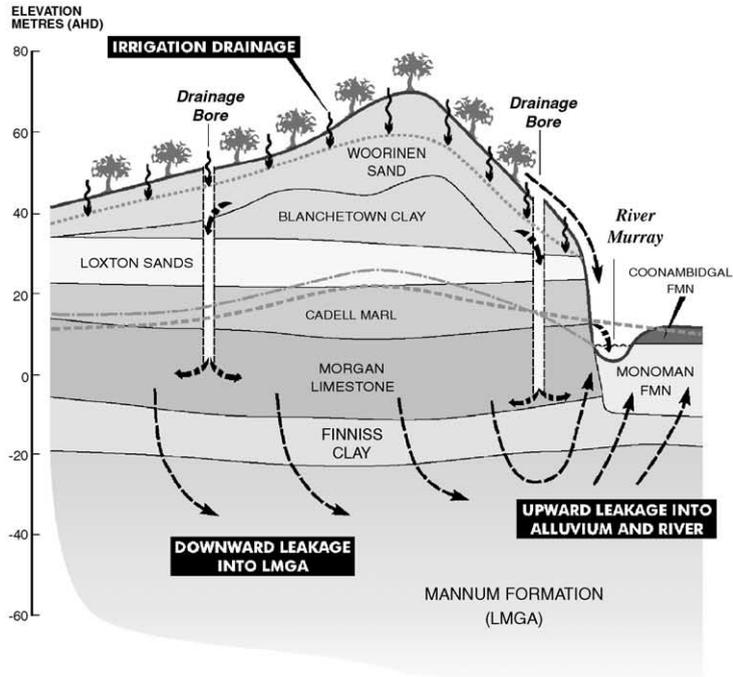
Fig. 1. Location map of the Murray Basin, and the Waikerie river stretch. Production bores of the Waikerie Phase I Salt Interception Scheme (SIS) are labelled. Production bores 4B, 5B and 6B had not commenced operation at the time of the survey.

erie irrigation area (Fig. 1), in which groundwater averages 20,000 mg/L total dissolved solids (TDS).

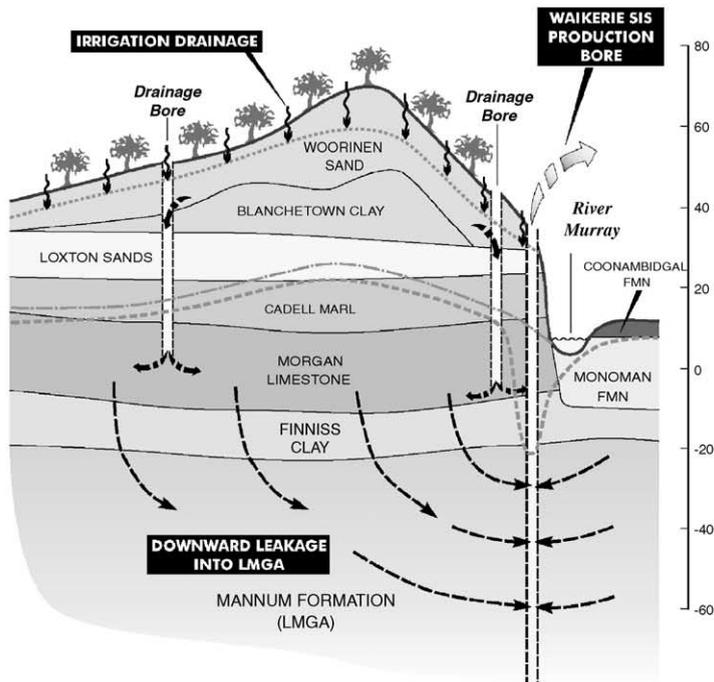
Stratigraphy of the survey region is shown in a cross-section across the River Murray at Waikerie in Fig. 2. The Murray Group Aquifer (MGA) system is the most intensively exploited regional aquifer in the Murray Basin (Evans and Kellet, 1989), deposited during a series of marine transgressions (Carter, 1985; Evans and Kellet, 1989). Blanchetown Clay and the Cadell Marl formations (Lukasik and James, 1998; Brown and Radke, 1989) are aquitards, allowing perched water-tables to form. When the perched water table reaches the root zone or the ground surface, waterlogging problems occur and crop quality is threatened.

An established local-scale treatment for water logging involves the drilling of drainage boreholes (Telfer and Watkins, 1991). These boreholes penetrate the Morgan limestone that constitutes the Upper MGA (as shown in Fig. 2a) allowing the perched water-table to drain. A regional impact is achieved with a high-density of drainage boreholes across an area: for example, in the Waikerie irrigation area approximately 280 drainage boreholes have been drilled in the last 100 years (Telfer and Watkins, 1991). Water flow through irrigation drainage boreholes have flushed and pressurised the Upper MGA, which in turn manifests high pressures in the Lower MGA. This hydraulic head increases the upward leakage of saline water through the Finnis aquitard,

**BEFORE IMPLEMENTATION OF WAIKERIE SIS**



**POST IMPLEMENTATION OF WAIKERIE SIS**



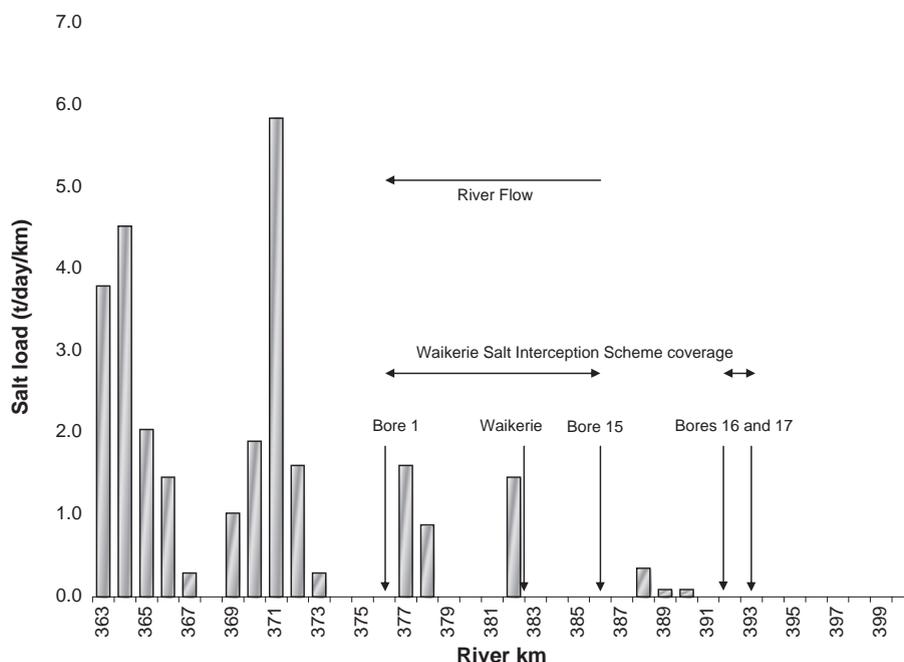


Fig. 3. Salt-loads per kilometre calculated from run-of-river surveys (downstream to left). Waikerie Salt Interception Scheme (SIS) production bores are labelled and can be located on Fig. 1. There is significantly less salt flowing into the river in the region of the SIS compared with further down stream, however small scale salt “hot spots” still exist. Spatial resolution of run-of-river surveys cannot be better than 1 km.

into the Monoman Formation and ultimately into the River Murray.

In an attempt to combat River Murray salinity, extensive salt-interception schemes (SIS) have been established (Lindsay and Barnett, 1989; Telfer and Watkins, 1991). These schemes consist of a series of boreholes that intersect the Lower MGA (as shown in Fig. 2b) close to the river, that are pumped to lower the excess hydraulic head (or pressure) of the aquifer, and hence reduce saline water flow into the River. Without upward hydraulic gradients by an over-pressurised Lower MGA, groundwater from the alluvial sediments leak down into the Lower MGA (Fig. 2b). Saline water from the boreholes is piped sometimes many kilometres to shallow lakes (either existing, or artificial) known as disposal basins, from which the water is left to evaporate. For the Waikerie SIS, the disposal basin is at Stockyard Plains, about 15 km from the river (Evans, 1989; Barrett et al., 2002).

For these schemes to effectively reduce river salinity, interception wells need to intersect the appropriate aquifer system, be spaced close enough to prevent upward leakage mid-way between wells (typically 500–1000 m) and be sufficiently close to the River to prevent saline inflow on both sides of the river. Borehole placement needs to take into account local variations in saline inflow that arise due to variations in irrigation activities and geological changes. However, many of these factors are poorly known, often based on pump tests with only one or two observation wells.

### 3. River Murray salt-load monitoring

Salt-load monitoring techniques are required along the River Murray to determine locations where saline-water accession is high, to aid in the development of salt interception schemes, and to assess the efficiency

Fig. 2. Stratigraphy of the survey region is shown in a cross-section across the River Murray at Waikerie. The upper figure shows regional hydrological flow and the seepage of saline fluids from the Lower Murray Group Aquifer (LMGA) into the River Murray. The lower figure illustrates the impact of the Waikerie Salt Interception Scheme (SIS), which lowers the water table in the vicinity of the River (figures courtesy of Australian Water Environments).

of existing mitigation schemes. Such monitoring techniques need to be fast and affordable so that the survey can be repeated for regular monitoring, and able to locate salt-load “hot-spots” to a scale of approximately 50 m.

The current practice is known as “run-of-river” (Porter, 1997; Telfer and Way, 2000). A run-of-river survey involves electrical conductivity (EC) measurement of surface river-water at 1 km intervals, which are converted to salinity. Measurements are repeated over five consecutive days. Water-bodies are tracked using flow data, and increases in salinity are plotted against river location. For example, for a river flow-rate of  $2 \text{ km day}^{-1}$ , the EC at a location is compared with the EC 2 km downstream on the following day. Locations where the water salinity increases significantly as it flows past can be interpreted as regions where groundwater inflow is high. Salinity increases are calculated as a salt-load in tonnes of salt-per-day-per-kilometre (t/d/km).

In-flowing saline-water is denser than the fresher river water, and changes in river depth and location of river bends are required for water to become well mixed (Telfer, 1989). As measurements are made at the river surface, anomalies are swept downstream before completely mixing with the water column. Salt-loads calculated from this method must therefore be corrected for downstream displacement. Run-of-river surveys are only feasible during stable river flow periods (Porter, 1997). Fig. 3 shows run-of-river salt-load data from the Waikerie river stretch.

Long-term measurements are made at salinity monitoring pontoons. Permanent mountings measure water EC (which is converted to salinity) every 30 min. Such pontoons are more widely spaced (10 km or more) than run-of-river sampling points, but can take data year round, showing seasonal variation in river salinity. They are also used to show year-to-year variation in the river salinity and to verify and calibrate run-of-river surveys (Porter, 1997; Vivian et al., 1998; Telfer and Way, 2000).

#### 4. Time domain electromagnetism

TEM survey methods measure the sub-surface electromagnetic response to a rapidly changing primary magnetic field generated by an electric current

flowing through a transmitter coil above the Earth. When the applied current is switched off rapidly (typically within  $1 \mu\text{s}$  for the NanoTEM system used in this survey) the primary magnetic fields collapses very quickly. An electric field associated with the time-varying magnetic field produces eddy currents in the Earth, that subsequently dissipate over time as energy is lost as a very small amount of heat. These eddy currents yield a transient secondary magnetic field. A receiver coil measures the time rate of change of this secondary magnetic field. In sediments with high-resistivity, eddy currents decay slowly and penetrate deeper; in low-resistivity ground, the currents decay more quickly (Reynolds, 1997).

To determine the resistivity of riverbed sediments, a floating TEM array was developed using Zonge Engineering and Research Organisation’s early-time NanoTEM system (Hatch et al., 2002), as shown in Fig. 4. A single-turn square transmitter loop of dimensions  $7.5 \times 7.5 \text{ m}$  was energised with a 3 A bipolar square-wave source (with equal on and off times and a repetition rate of 32 Hz). A central  $2.5 \times 2.5 \text{ m}$  single-turn square receiver loop measured the transient secondary magnetic field. These loop sizes are smaller than typical TEM systems, however faster turn off times help to prevent typical problems that small loops can cause. The frame was constructed from timber and strengthened using fibre-tape and diagonal ropes and secured using nylon bolts. Floatation was achieved using tyre inner tubes. The frame was towed 10 m behind the boat. Towing speed was kept as low as possible, at about  $3\text{--}5 \text{ km hr}^{-1}$ , to minimise strain and vibration on the array.

Each TEM response was determined from sixty four soundings stacked at 32 Hz and sampled at intervals of  $1.6 \mu\text{s}$ , with the final time window (channel 31) recorded at 2.5 ms. For most soundings, only windows 4 to 15 (of the 31 recorded) were used; earlier and later windows had low signal-to-noise ratios. These windows correspond to  $5.3 \mu\text{s}$  and depths of 4 to 20 m for window 4 to  $64 \mu\text{s}$  and depths of 20 to 70 m for window 15. Depths are estimates from current diffusion for a range of 2 to  $50 \Omega\text{m}$  ground resistivity (Nabighian and Macnae, 1991). Actual signal depth at the 15th window (as indicated by the best fitting models) was in the range 15 to 25 m.

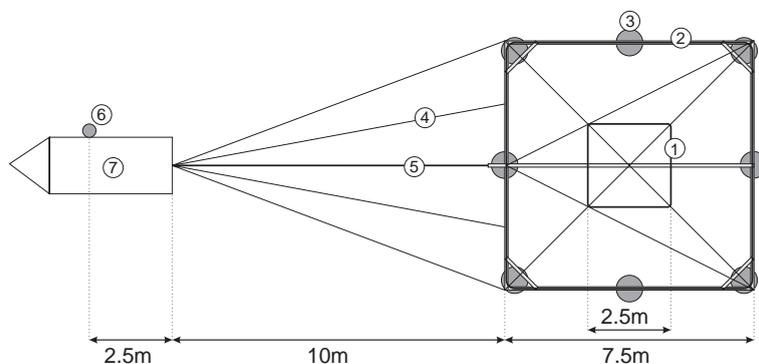


Fig. 4. Floating deployment of the Zonge NanoTEM. Labelled are (1) receiver loop, (2) transmitter loop, (3) tyre inner tubes, (4) tow ropes and diagonal support ropes, (5) 10 m spacer and cable housing, (6) GPS and echo sounder units and (7) the boat.

Sensitivity of the TEM method to detecting resistivity of riverbed sediments from a floating transmitter was investigated. One-dimensional modelling (assumes infinite horizontal layers) and inversion was carried out using the program STEMINV (MacInnes and Raymond, 2001), which simultaneously minimises model roughness and misfit criteria. Smoothness of the final model is dependant on the user-controlled smoothness constraint that trades-off model roughness with misfit. For sensitivity analysis, single-station models were defined using configurations of 16  $\Omega\text{m}$  river water, 50  $\Omega\text{m}$  freshwater-saturated sediments (assuming porosity between 50% and 60%), a conservative conductive value at 10  $\Omega\text{m}$  (for example 6000  $\text{mg L}^{-1}$  saline water in sand) and a very conductive value of 0.5  $\Omega\text{m}$  (equivalent to 27,000  $\text{mg L}^{-1}$  saline water in a sediment with some conductive clay content). These values were chosen based on Archie's Law estimates of conductivity for the given porosity and salinities (Archie, 1942).

Where 5 m of river water overlies very conductive sediments and no smoothing is applied to the inversion (Fig. 5a) results indicate that the location of the boundary corresponds approximately with the inflection point on the inverted model. This figure also shows that the inverted model tends to oscillate about the conductive value with depth. The depth instability can be reduced by increasing the smoothness constraint in the inversion (Fig. 5b), but reduces the resolution of the boundary. If gradational boundaries are encountered, the smooth

inversion is much more successful at reproducing the original model (Fig. 5c). While a smooth boundary is not expected at the River water–sediment interface, most other resistivity changes encountered in this project are likely to be gradual. Fig. 5d shows that when a 5 m thick resistive layer occurs at a depth of 5 m, the resistivity value is significantly underestimated. In contrast, the resistivity value of a conductive layer of the same thickness is well determined, except that resistivity is too high both above and below the conductive layer (Fig. 5e). In both the resistive layer and the conductive layer case, smoothing results in a broader region than the true layer thickness, with a half-width of almost 10 m.

From this study, a list of guidelines for successful interpretation of STEMINV smooth inversion has been established:

1. Smoothing criteria provide better constraints on lower layer resistivities (by stabilising the oscillations of the model with depth);
2. Where sharp or rapid resistivity changes are expected (e.g., water–sediment boundary) the smoothing criterion reduces the resolution of the boundary;
3. Where gradual resistivity changes are expected (e.g., most hydrology related changes) the estimate of resistivity can be very close to the true resistivity, regardless of smoothing criteria; and
4. The resistivity of conductive layers is more accurately determined than that of resistive layers (which are underestimated).

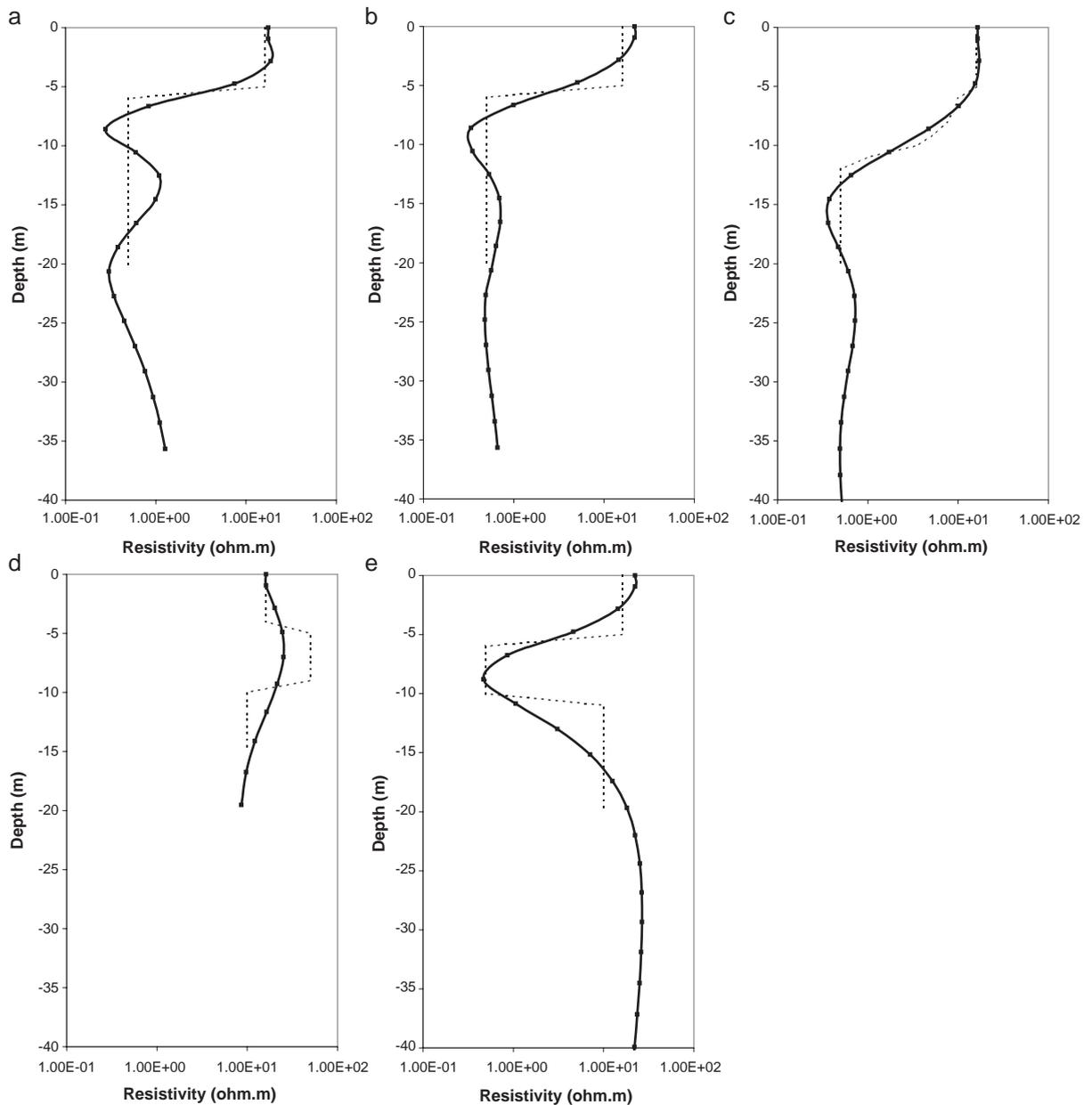


Fig. 5. Sensitivity tests for the NanoTEM system used in the surveys. In each figure, the dashed lines are the models from which TEM responses are calculated. Solid lines represent the inversions of the modelled responses using the program STEMINV (MacInnes and Raymond, 2001). (a) A step down in resistivity at 5 m depth from 16 to 0.5  $\Omega\text{m}$  without smoothing (smoothing factor of 1.0); (b) As for (a) but with a smoothing factor of 3.0; (c) A gradual or smooth change in resistivity is fit closely by the inversion; (d) A 5 m thick layer increase in resistivity from 16 to 50  $\Omega\text{m}$  is poorly detected by the inversion; (e) By contrast, the inversion of a 5 m thick decrease in resistivity from 16 to 0.5  $\Omega\text{m}$  yields the correct resistivity, but is much broader than the layer thickness.

Ideally, a graduated smoothness function should be applied, so that in the top 10–15 m, where a sharp resistivity boundary is expected, minimal smoothing can be applied, while at depth, where resistivity should vary more slowly, smoothing can be used to stabilise the model.

## 5. River TEM survey

River TEM surveys were undertaken in two separate experiments. In August 2002, 9 km were acquired between river km 376 and 386 (survey line 1 in Fig. 6), along a stretch of the river that is currently pumped by the Waikerie salt-interception scheme (Boreholes 4 to 15). A sub-section of this survey (river km 377 to 381) was measured again in December as a test of repeatability. In the latter survey, four survey lines were recorded (lines 4 to 7), each separated by approximately 50 m, to enable changes in the response across the river to be observed. During the December survey, two survey

lines (survey lines 2 and 3) along an “unpumped” river stretch (river km 368 to 376) were also recorded. These surveys total 40 km of data, which were acquired in a total of approximately 15 h. GPS locations were recorded with a Garmin handheld unit located on the boat, while water depth was recorded at one-minute intervals from an echo sounder. Water resistivity was also recorded from the boat.

Resistivity inversions were carried out using the STEMINV program. Fig. 7 shows raw data and the best fit smooth model that was determined for one station. These smooth 2D models were stitched and horizontally smoothed then gridded and displayed using the Geosoft Corporation’s Oasis Montaj package giving both two and three-dimensional figures. Fig. 8 shows four parallel profiles (survey lines 4 to 7 in Fig. 6) arranged with the northern most line at the top. Resistive anomalies produced by pumping from SIS boreholes can be identified on each of the survey lines, but the effect is greatest on the lines closest to the southern river bank. The conductive region be-

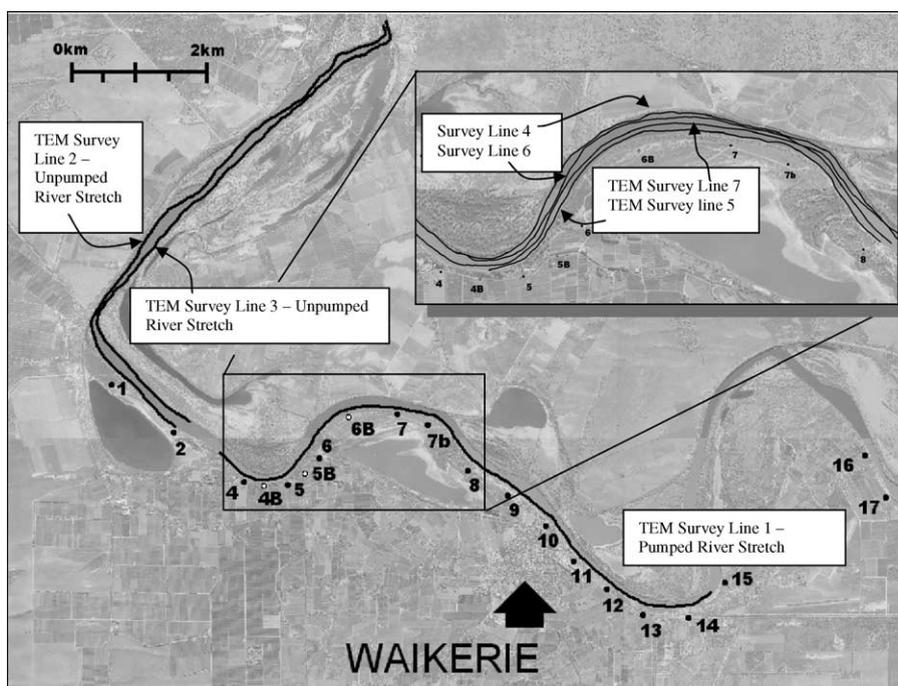


Fig. 6. Location map of survey lines. Lines 2 and 3 are in the unpumped survey section, while lines 1 and 4 to 7 are in the pumped survey section. Waikerie Salt Interception Scheme production bores are labelled.

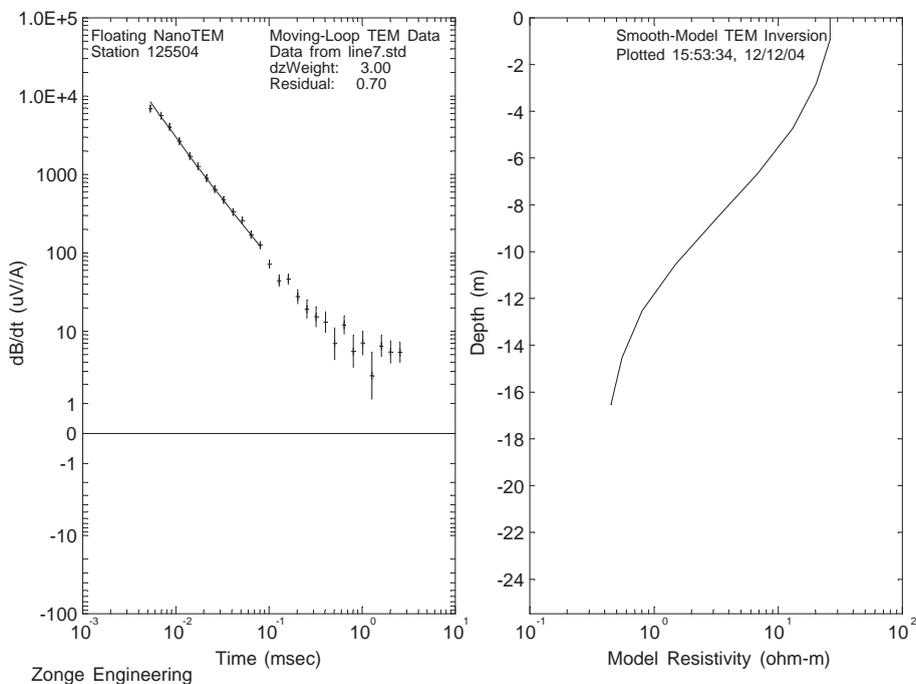


Fig. 7. Raw data (crosses) from a single survey station. Time windows 1 to 3 were discarded before plotting. Time windows 4 to 15 have been used in this inversion, with the smooth model shown on the right and the closeness of fit to the data visible on the left.

tween SIS boreholes 6 and 7 is more extensive, and closer to the river bed in the northern most lines.

A plan view of the resistivity of the top 5 m of alluvial sediment is shown in Fig. 9, which was produced by averaging the resistivity of the top 5 m of alluvial sediment at each data station, and creating a horizontal, minimum-curvature grid of the data. The major conductive anomalies occur 1–2 km upstream of run-of-river salt-load highs, because denser saline water is not detected at the surface by the run-of-river technique until changes in river depth or bends in the river cause mixing to occur (Telfer, 1989). The most significant conductive regions in this figure correspond with deep points in the river, in agreement with Telfer (1989) who used theoretical models to calculate that river trenches provide 40–60% higher accession rates of aquifer water to the river than other locations.

The river stretch not pumped by the Salt Interception Scheme is shown in Fig. 10 (survey lines 2 and 3) with the western most line on top (see Fig. 6 for survey line locations). Data show conductive features beneath the riverbed on the southern ends

of the survey lines (at the left of the figure), and downstream of borehole T5 toward the northern end of the survey lines (towards the right of the figure). Differences between the two sides of the river can be seen, and in particular, a low resistivity feature 500 m downstream of borehole T5 appears more prominently on the western side than on the eastern side of the river.

## 6. Interpretation of conductivity variations

Between boreholes 10 and 15 in Fig. 6 the Glenforslan Formation, (a sublithified to unlithified limestone) outcrops as the section changes from flood plain to highlands in this area. This region is significantly more resistive in the models than other sections of the survey, suggesting some relationship between the Glenforslan Formation and resistive conditions. However, in other locations resistivity changes over a much smaller spatial wavelength than is likely for geological variations alone. For example, in Fig. 9, localised resistive

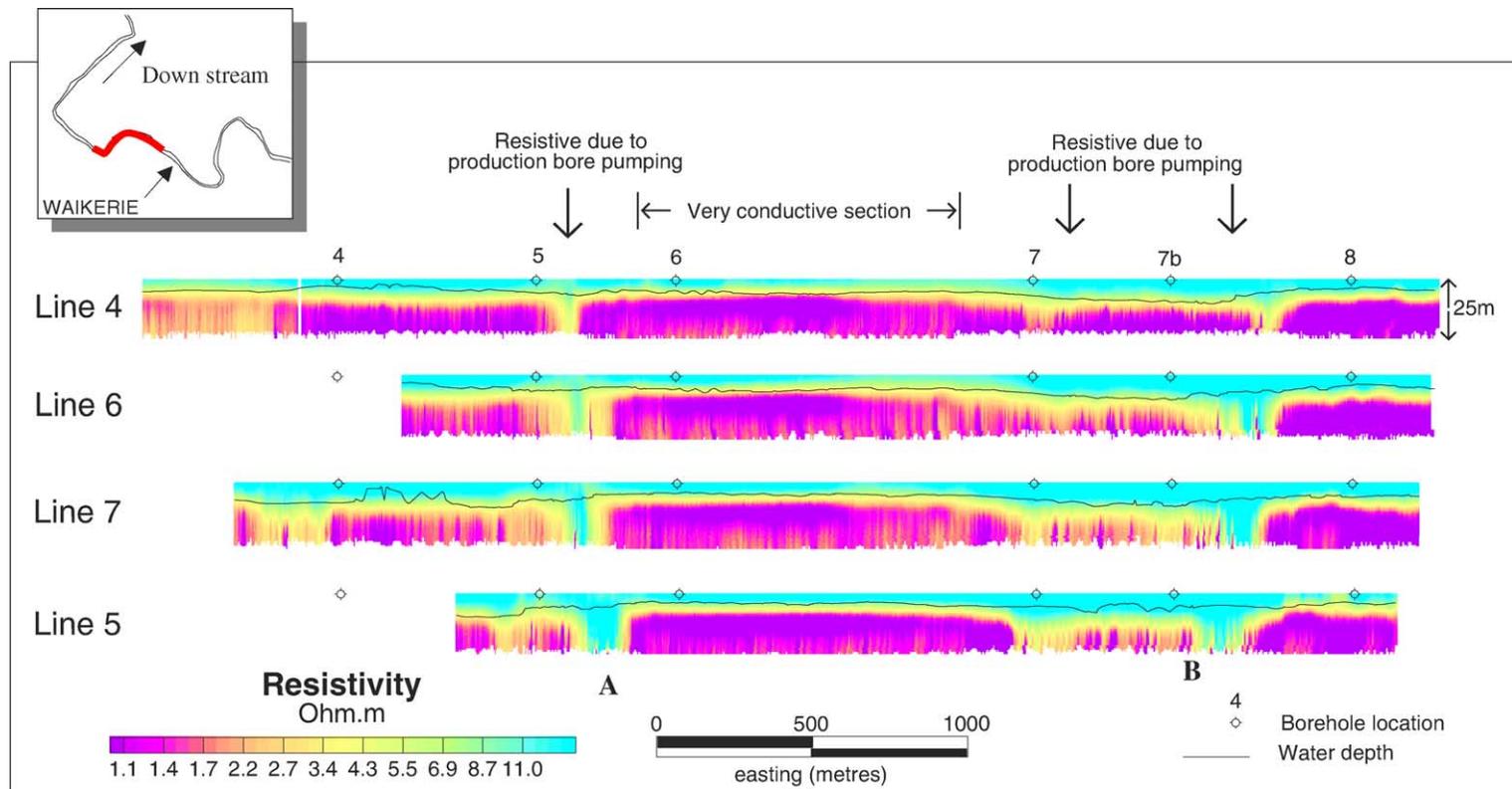


Fig. 8. Four parallel profiles (survey lines 4 to 7 in Fig. 6) arranged with the northern most line at the top. The sub-horizontal black line shows the water depth. Resistive anomalies produced by pumping from production bores can be identified on each of the survey lines, but the effect is greatest on the lines closest to the southern river bank. The conductive region between production bores 6 and 7 is more extensive, and closer to the river bed in the northern most lines.

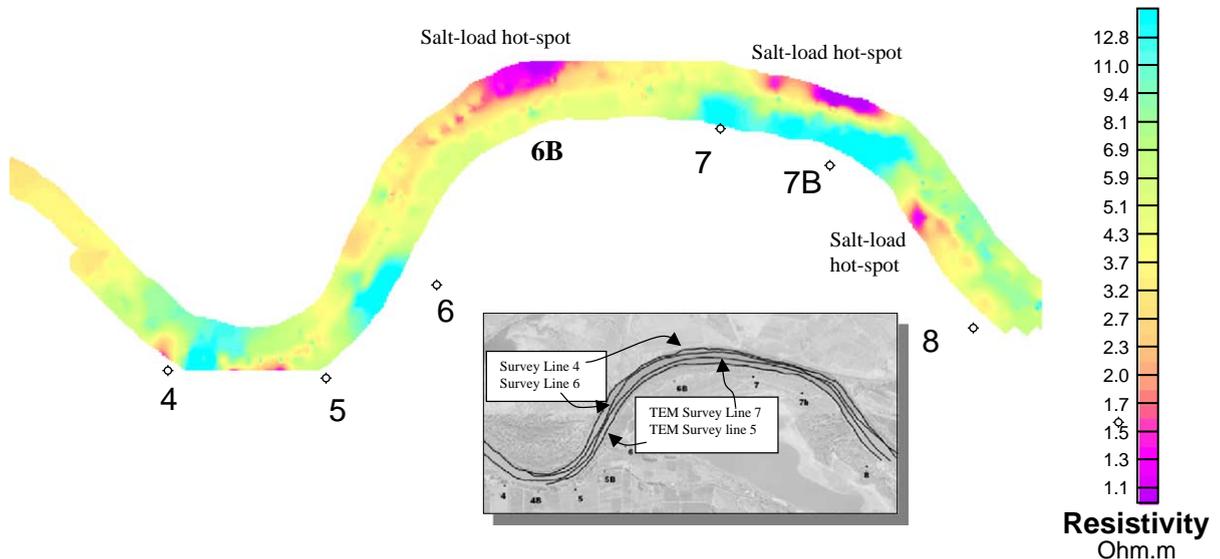


Fig. 9. A plan view of the average resistivity of the top 5 m of alluvial sediment. At time of survey, planned borehole 6B was not operational.

anomalies around the Waikerie SIS boreholes 4, 6, 7 and 7B suggest a hydrogeological response from pumping, as fresh river-water replaces saline-water in the alluvium. Thus, we associate resistivity variations more with hydrology than with lithology in this section.

Variation in clay content may also affect sediment resistivity beneath the river. If clay-conduction is sufficiently efficient (compared to non-clay sediments), low-resistivity anomalies may be due to the occurrence of clay. However, clay layers have an important impact on the resistivity through another mechanism; low permeable clays act as aquitards and confining layers. Such hydrological control exhibited by clay may be more significant than clay-conduction if fluid conductivity is highly variable. Sediment sampling is currently underway to ground truth the measurements in terms of clay content.

The most significant conductive anomaly in the top 5 m of alluvium in the pumped section of the survey (Fig. 9) was midway between SIS boreholes 6 and 7. Borehole 6B, which had not commenced operation at the time of the survey, is located in the same region. A geological log from the drilling of production borehole 6B revealed that the Finnis Formation was not present. The Finnis Formation is known to be a confining layer for the Lower MGA, and consequently increased saline-water in-

flow should be anticipated where this confining layer is absent. Absence of the Finnis Formation at borehole 6B is evidence that the anomaly is a salt-load hot-spot, and that hydrology is more important than clay-conduction in these models. Other anomalies that were identified as potential hot-spots are located between boreholes 7 and 7B and between boreholes 7B and 8 (Fig. 9), which were smaller and less conductive than the anomaly between boreholes 6 and 7.

Some SIS boreholes show little or no resistivity correlations (for example at boreholes 5, 8 and 9 in Fig. 9). Efficient pumping of SIS boreholes prevents saline water flowing into the river, but not enough to induce significant drawdown of fresh water, with the result that neither conductive nor resistive anomalies are seen. Alternatively, the permeability of aquitard units above the pumping depth may be low in these locations, preventing draw-down of fresh water into the sediments; and, near surface regions with high permeability may result in lateral fluid flow from locations offset from the borehole location.

Most of the unpumped survey line lies at the edge of the flood plain. There are no known significant geological changes along this section of the river, but the survey line is in close proximity to the Glenforlan Formation. The section is not as resistive as the region with similar geology in the pumped section

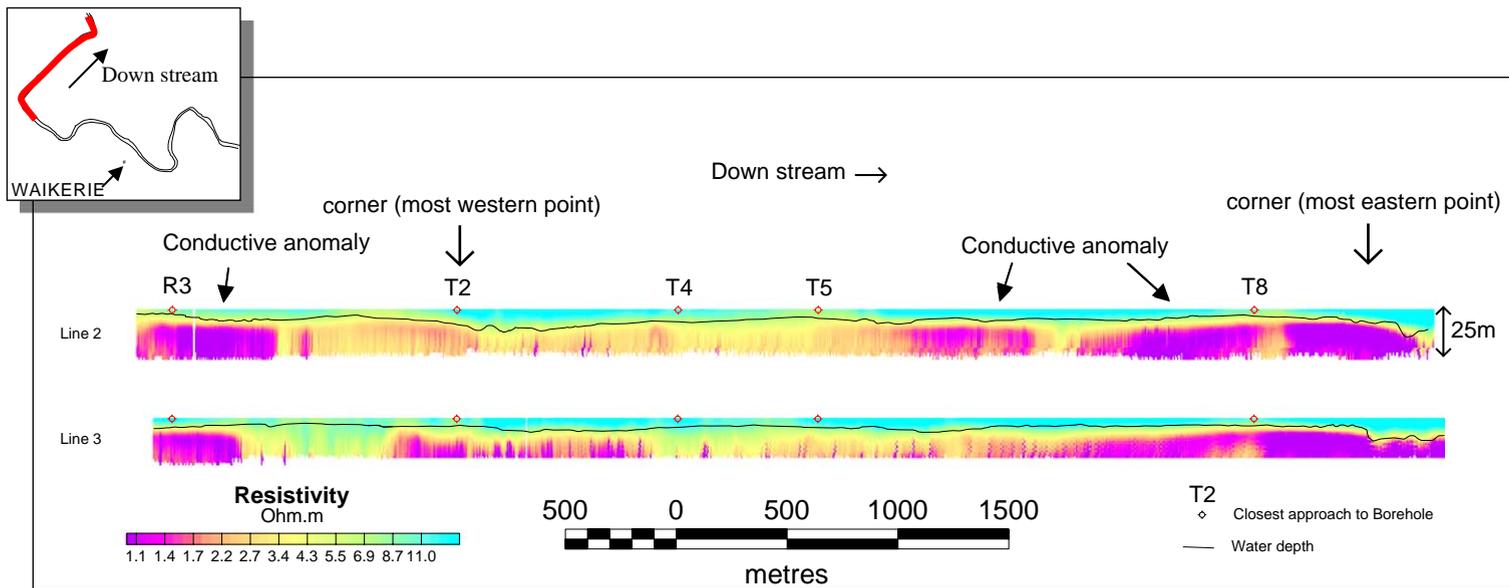


Fig. 10. The river stretch not pumped by the Salt Interception Scheme (survey lines 2 and 3) with the western most line on top (see Fig. 6 for survey line locations).

(between boreholes 10 and 15), but could become very resistive in response to pumping if the Glenforlan Formation acts as a permeable conduit for fresh-water drawdown.

## 7. Archie's law interpretation of salt load

Archie's law (Archie, 1942) is an empirical law that relates bulk resistivity to porosity, saturation and fluid resistivity:

$$\rho_b = \frac{\rho_w}{a\phi^m S^n} \quad (1)$$

where  $\rho_b$  is the bulk resistivity of the formation,  $\rho_w$  is the resistivity of the contained fluids,  $\phi$  is the formation fractional porosity,  $S$  is the formation fractional saturation and  $a$ ,  $m$  and  $n$  are empirical constants. In the present study  $a$  and  $m$  were set to 1 and 2, respectively, while the value of  $n$  was irrelevant because saturation  $S$  was 1 as the sediments were 100% saturated. The porosity was taken as 55% based on typical sediment compactions (Telfer pers. comm. 2003).

Table 1 shows an interpretation of significant resistivity features from the cross-sections in Fig. 8 from Archie's Law. In each case any possible contribution of clay has been neglected: clay would result in an estimate of fluid resistivity that is higher than the following estimates. Inversions using STEMINV are likely to underestimate the resistivity in more resistive areas, and thus the true fluid salinities are probably lower than estimated in Table 1. Salinities at resistive anomalies (e.g., around the SIS boreholes) may be closer to river water salinity (330 mg/L TDS) than the estimated 1200 mg/L TDS.

Table 1  
Observed resistivity measurements, inferred formation fluid resistivity based on Archie's Law and the interpreted salinities

Region	Bulk sediment resistivity $\Omega\text{m}$	Formation fluid resistivity $\Omega\text{m}$	Salinity mg/L TDS
Major hot-spots	1	0.3	18,200
Deep conductive features	0.5	0.15	36,400
Background resistivity	5	1.5	3600
Resistive anomalies at SIS boreholes	30–50	3–4.5	360–600

## 8. Conclusions

Our adaptation of an existing TEM system for deployment over water allowed us to measure the conductivity of the top 5 m of alluvial sediments on a stretch of the River Murray, South Australia. Locations of significantly high saline water accession were inferred from high conductivity anomalies. Our surveys measured two distinct regions; one under the influence of a Salt Interception Scheme, where anomalies were seen as targets for scheme improvement, the other in a region not influenced by such schemes. We have demonstrated the application of this system to mapping of conductive features beneath the river for the purpose of identifying problem saline water accession areas, however ground truthing of the results has not been conducted.

Sediment sampling along a location with a near-surface conductive anomaly could provide a degree of ground truthing for the TEM technique. Measurement of porosity, bulk resistivity, fluid resistivity and clay content of collected samples would be required. Data acquired from such a sampling program could be compared to fluid salinity estimates made in this project using Archie's Law. Ground truthing could also be achieved with a follow up survey after borehole 6B of the Waikerie SIS commences pumping. If the described conductive anomaly (major salt-load in Fig. 9) is due to hydrology, then pumping from borehole 6B will reduce the size of the anomaly. This is considered the most important repeat survey requirement for further assessment of the TEM technique for salt-load monitoring.

Since this project was undertaken, the Technique has been employed on some large scale mapping of River Murray salinity problem areas, with over 800 line kilometres being measured.

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