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On the effects of irrigation and urbanisation on the annual range of monthly-mean temperatures

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With 2 Figures

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Summary

Weather station data indicate that large-scale irrigation of an arid region in southeastern Australia decreased the annual range of monthly-mean temperatures by 1–2 K. Such decrease is consistent with an observed increase in dewpoint, by about 1 K. Urbanisation may reduce the annual range but generally increases it.

1. Introduction

The relationship between a) the annual range of monthly-mean temperatures on land and b) geography (latitude, distance from the upwind shore, topography) was explored by means of global weather station data in Geerts (2002) (abbreviated as G02). That study is the basis of the present note, which examines the effect of permanent, man-made changes in land use or land cover on the annual temperature range (ATR). The main resources used here are a) the International Station Meteorological Climate Summary, version 4.0, issued on a CD-ROM by the US National Climate Data Center in 1996 (referred to as ISMCS), and b) the historical record of monthly-mean data of all weather stations in Victoria and New South Wales, Australia, obtained from the Australian Bureau of Meteorology (referred to as BoM).

2. Adjusting the annual range for differences of latitude and distance inland

The effect of contrasting land use on ATR is here examined in the case of the Murrumbidgee-Coleambally-Murray Irrigation Areas (MIA), around 35° S in south-east Australia. These are situated on a semi-arid plain at least 220 km from the ocean, which is found to the south-west, south and south-east. The plain's elevation ranges between 50–250 m above sea level, with higher elevations at the eastern margin. Annual rainfalls vary from about 300 mm in the north-west of the MIA to 410 mm in the southeast, with a slight maximum in the winter months. The warmest and coldest months are January and July, respectively.

The ATR is controlled mainly by latitude and distance inland, so to analyse the influence of regional land use on ATR, one needs to remove their effect first. The procedure, obtained in G02, is summarized here. As regards latitude, station data from the southern continents indicate that the meridional ATR gradient over land is close to linear at about 0.3 K/degree latitude. As regards the effect of the fetch from the coast, the ATR increases inland, but its rate depends on the presence of topographic barriers and on the wind direction. The latter varies over the MIA, being

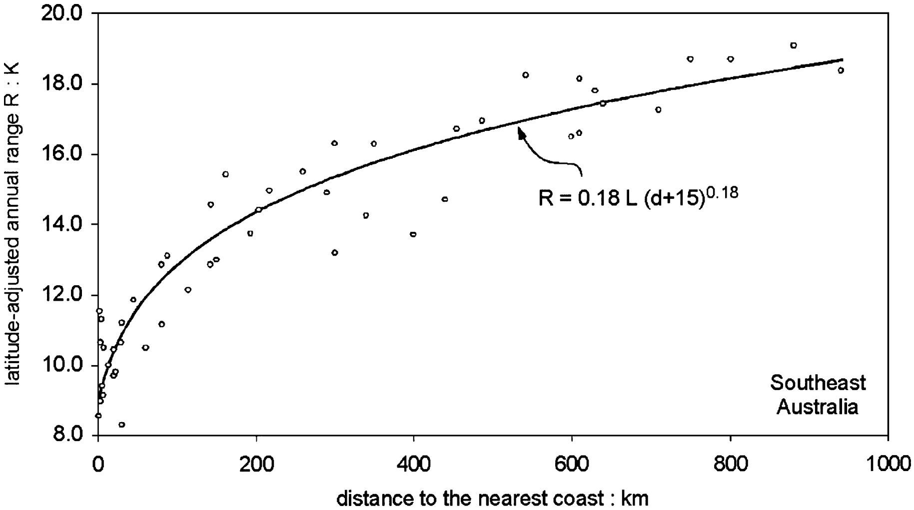

Fig. 1. The variation with distance to the nearest coast of the ATR (after adjustment to a common latitude of 30° S) for 50 stations in southeast Australia, between 26–37.5° S and 133–153° E (ISMCS data)

most commonly westerly in winter and southerly to easterly in summer. In G02, the distance is computed from the nearest coast to the east or west, depending on the prevailing zonal wind direction. Here the distance inland is computed simply as the distance to the closest coast, because there is ocean in all three directions (west, south, and east) at roughly the same distance. The influence of the Dividing Range on the ATR has not been removed. This topographic barrier is 1–2 km high and is located to the east of the MIA.

The observed annual range of monthly-mean temperature at each of 50 stations in the southeast quadrant of Australia is 'adjusted' to a reference latitude of 35° S, assuming 0.3 K/degree as the ATR gradient with latitude L (degree). A regression analysis of the adjusted ATR R (K), latitude L (degree), against distance from these stations to the closest coast d (km) yields the following relationship (Fig. 1):

$$R = 0.18\ L(d + 15)^{0.18}\ K \qquad (1)$$

The form of this equation is based on previous work (G02), though the coefficients are slightly different because the equations in G02 are based on global station data, and because the distance d in G02 is defined as the distance to the coast in an east or west direction, depending on the prevailing zonal wind.

Equation 1 is used to remove the effect of distance inland on the ATR values observed in the MIA region by adjusting the observations to their equivalents at a common distance of 350 km for the nearest coast. For instance, Ivanhoe (32.9° S) has an ATR of 16.4 K and is 557 km from the nearest shore. The equivalent annual range at 35° S is 17.0 K [i.e. $16.4 + 0.3\ (35–32.9)$], and if the ocean were 350 km away the range would be 15.7 K [i.e. $17.0\{(350 + 15)/(557 + 15)\}^{0.18}$]. The latter value is referred to as the adjusted ATR for Ivanhoe.

3. The effect of regional irrigation on annual range

The spatial variation of ATR, adjusted for the dominant effects of latitude and distance inland, is now examined about the MIA, in a region smaller than that used for Fig. 1. The Australian Bureau of Meteorology maintains a dense network of weather and climate stations in southeast Australia and monthly-mean temperature data from 28 of these stations for the period 1968–1996 were used to study the effect of irrigation within the MIA. The irrigated areas (Fig. 2) are not contiguous but the total MIA domain is of mesoscale dimensions (13.824 km^2). The MIA was largely established by 1968. Its extent has changed little since then, except for the smaller Coleambally area, which expanded in the 1970s and later shrunk somewhat.

The adjusted ATR increases from west to east across the MIA (Fig. 2), mainly at the eastern margin of the map, perhaps because several stations there are located in valleys, and because the eastern margin is in the lee of the Dividing Range in summer. Both of those factors (a terrain concavity and the blockage of low-level airflow by a mountain range) enhance the ATR (G02).

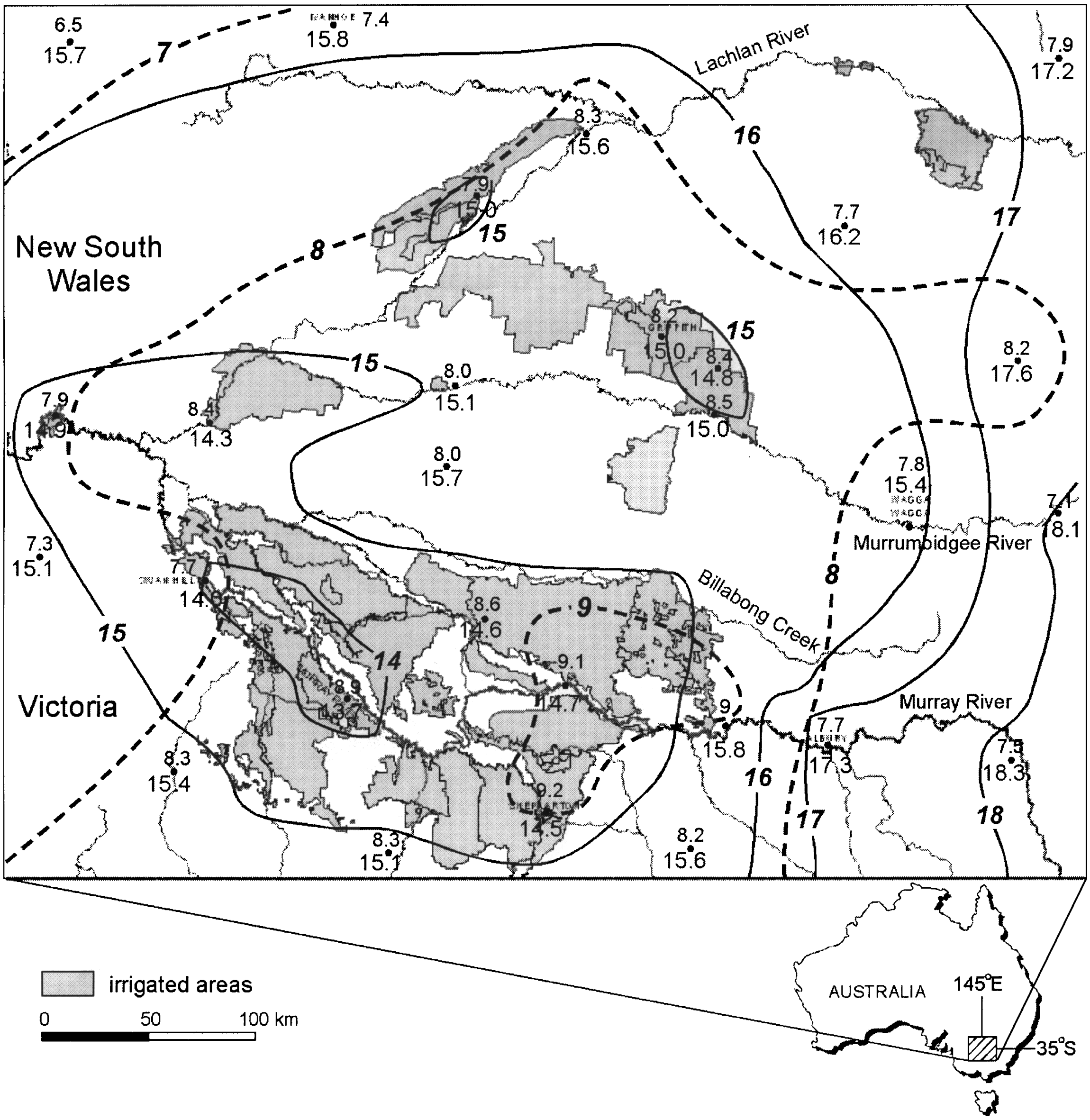

Fig. 2. Latitude- and fetch-adjusted ATR (K, shown below the station positions), and annual-mean dewpoint (°C, above station position) at 28 stations about the Murrumbidgee-Coleambally-Murray Irrigation Areas of Australia. The climate data are 1968–1996 BoM averages. Isopleths of dewpoint (dashed lines) and ATR (solid lines) have been added by eye

Superimposed on the west-to-east ATR gradient is a minimum in ATR over the irrigated areas. A deficit in adjusted ATR of 1–2 K is found over the MIA. Unadjusted ATR values (not shown) do not display this minimum, mainly because of the strong dependence of ATR on distance from shore: unadjusted ATR values merely show a local reduction of the general gradient (approximated by Eq. 1) over the MIA. The adjusted ATR deficit is largely due to the damping of January (summer) temperatures within the MIA (Table 1). Adjusted temperatures within and around the MIA differ by less than 0.2 K on average in July, when little irrigation occurs.

The ATR deficit at the core of the MIA is unlikely to be an artifact of the adjustments for latitude and fetch. This is indicated by the expected matching pattern of annual-mean dewpoints, also shown in Fig. 2. The annual-mean dewpoint temperature is computed from daily averages, based on 9 am and 3 pm observations. The dewpoint tends to fall from the east to increasingly arid land in the northwest. Superimposed on this trend is a maximum in dewpoint

Table 1. Average values of temperature, humidity, and precipitation for the 12 stations within the MIA (in Australia) and the 16 stations outside the MIA shown in Fig. 2. The data are based on the 1968–1996 period

Variable	Within the MIA	Outside the MIA
Annual mean temperature (°C)[1]	16.8	17.3
January mean temperature (°C)[2]	24.0	25.4
July mean temperature (°C)[2]	9.1	9.2
ATR (K)[2]	14.8	16.2
January mean dewpoint (°C)[3]	12.3	10.8
July mean dewpoint (°C)[3]	5.1	4.7
January mean precipitation (mm)	30.1	38.3
July mean precipitation (mm)	36.2	46.0

[1] Corrected to sea level and to a reference latitude of 35° S, as in Linacre and Geerts (2002).
[2] Adjusted to a reference latitude of 35° S and to a distance of 350 km from the nearest shore, as discussed in Section 2. Half of the ATR adjustment is added to the January temperature, and half is subtracted from the July temperature.
[3] Average of 9 am and 3 pm data.

over the MIA, mainly its eastern margin. A dewpoint excess of ~ 1 K (i.e. a mixing ratio of ~ 0.5 g/kg) is found in the MIA, ranging from 1.5 K in January (summer) to 0.4 K in July (Table 1). The extra atmospheric humidity does not translate into a regional rainfall maximum. In fact the stations within the MIA receive about 20% less rain than those outside the MIA, both in January and July (Table 1).

Additional evidence on the effect of changing land use on the ATR should arise from a comparison of seasonal temperatures before and after the development of the irrigation areas. Unfortunately there were only six stations in the region before the establishment of MIA. Records at these stations commenced before 1900, except at Griffith, where they commenced in 1914. Of these six, only those at Griffith and Deniliquin are located within the MIA. The average ATR at these two stations (adjusted for latitude and inland fetch) was 0.3 K smaller than that at four nearby stations outside the MIA during 1900–1950, but 1.1 K smaller during 1968–1996 (BoM data). This increase in ATR deficit within the MIA (at Griffith and Deniliquin) between the periods 1900–1950 and 1968–1996 was due to an ATR decrease within the MIA by 0.9 K, not to any ATR increase at the four surrounding stations: their ranges remained the same (i.e. a reduction of only 0.1 K). Moreover the ATR decrease at Griffith and Deniliquin was essentially due to January cooling (-0.8 K), not to July warming ($+0.2$ K). This evidence suggests that the ATR deficit in the irrigation area was insignificant before irrigation commenced, and that it is truly due to changes in land use.

4. The effect of urbanisation on annual range

Monthly-mean data from the climatological ISMCS record are used to compare the ATR in five American cities with those in their environs. The ATR in downtown St. Louis is 27.8 K, which is 1.2 K more than the average ATR at three rural stations nearby, two to the east and one to the west. The ATR in the Baltimore central business district is about 2.5 K more than that at an adjacent rural site (Landsberg, 1981). The ATR in downtown Atlanta and Phoenix is marginally more than at two stations outside each city (0.4 K and 0.6 K respectively). But the ATR in downtown Chicago is 1.7 K less than the average ATR of three stations within 53 km to the northwest, west and south (ISMCS data).

These five examples suggest that the ATR is indeed affected by urbanisation, though the effect may be obscured by other influences, such as proximity to water or complex terrain. Urbanisation enhances the ATR in topographically uniform areas such as around St. Louis. The urban ATR excess in the four cities mentioned above (St. Louis, Baltimore, Atlanta, and Phoenix) is mostly due to warmer summers: the city centers are, on average, 1.3 K warmer than their surroundings in July, instead of 0.2 K in January. The ATR deficit of 1.7 K in downtown Chicago is anomalous and is caused by cool breezes from Lake Michigan in summer: the July-mean temperature in downtown Chicago is 2.0 K below that of the three adjacent stations. Perhaps the absence of a substantial ATR excess at Atlanta is due to its high density of deciduous trees: the July mean temperature in central Atlanta is about the same as that in the surrounding (the excess is 0.3 K).

It may be argued that the urban ATR excess in the above and other examples is due to some topographic coincidence rather than to the change in land cover, i.e. urbanisation itself. Even subtle variations in topography can have a significant influence: the ATR in Swiss valleys is about 5 K larger than that on hills at the same elevation (G02). In fact many cities are located in valleys. However, the observed seasonal asymmetry of increased urban heating as a city grows confirms the effect of urbanisation on ATR. The rate of warming of summers in three large American cities (Washington DC, Cleveland and Boston) over the past century has been over twice that of the rise of winter temperatures (Changnon, 1992). As a result, these cities experienced an increase of ATR by 1.0–1.8 K in the last century, while the topography remained unchanged.

5. Discussion

The examples given in this note illustrate that land use and land cover affect the ATR. The observed variations in ATR can largely be understood in terms of observed differences in diurnal temperature range (DTR). Natural and manmade differences in land surface conditions have been shown to cause DTR differences (Dyer and Crawford, 1965; Landsberg, 1981; Bonan, 2001). The reason is that either the net radiation at the surface differs (due to differing albedo), and/or the Bowen ratio of surface fluxes varies. The irrigation of normally arid land increases evaporation from surfaces with sufficient soil moisture and lush vegetation. This reduces the ground temperature during the day, and therefore dampens the daily maximum air temperature (Dyer and Crawford, 1965).

The DTR may be reduced further by atmospheric changes in large irrigation areas. The enhanced evaporation increases the absolute humidity in the atmospheric boundary layer, as observed in the MIA. Increased humidity decreases nocturnal heat loss, because water vapor is a greenhouse gas. Increased humidity may also increase cloudiness, which decreases both daytime heating and nighttime cooling. Numerical simulations suggest that these atmospheric changes are insignificant in small areas, and that their significance increases with the size of the irrigation area and the aridity of the environment (Stohlgren et al., 1998). These atmospheric responses have been documented under natural forcing, for instance unusually heavy rainfall in a mesoscale region may enhance cloudiness and further rainfall over the following few days, especially in summer, through changes in the surface energy balance ('precipitation recycling', Eltahir and Bras, 1996).

The reduction in DTR is especially remarkable near large irrigated crop areas in an arid environment, as evident from station data (Gallo et al., 1996; Dai et al., 1999) and from numerical simulations (Bonan, 1997; Stohlgren et al., 1998). Both modeling and observational evidence indicates that the DTR reduction is mainly the result of lower daytime maxima in the growing season, not higher nighttime minima (Bonan, 2001). This implies lower monthly-mean temperatures in summer over irrigation areas, as observed over the MIA (Table 1). The DTR and mean temperature are hardly affected in winter when irrigation activities cease or are reduced, as is the case in the Great Plains of North America (Bonan, 2001) and in the MIA.

In short, the same processes that explain a reduction of DTR due to regional irrigation are responsible for the reduction of ATR as observed in the MIA. The lower daytime maxima in the growing season also imply a lower mean temperature during the warmest month and hence a smaller ATR.

The excess dewpoint of about 1.5 K in the MIA in summer (Table 1) corresponds to an excess specific humidity of about 0.9 g/kg. If the latent heat required to evaporate this water is balanced by the cooling of the adjacent air, then this humidity excess implies a cooling of 2.2 K. This value is not much larger than the adjusted January temperature deficit in the MIA (Table 1), confirming that evaporative cooling is the main process explaining the reduced ATR over the irrigated area.

The MIA in Australia is ideally suited to examine the effect of land use on ATR because of its inland location and flatness. An increase in specific humidity of up to 0.3 g/kg in summer has been documented between 1964–1995 in the lower troposphere above the coastal region of Israel, an area that is smaller than the MIA but with a similar expansion of irrigation activities (Ben-Gai et al., 2001). During the same period neither the temperature nor the ATR changed significantly over the Israeli coastal region. Though the distribution of ATR in the heavily irrigated San Joaquin and Imperial Valleys in Southern California is well-documented by the historical climatological network (e.g. the ISMCS), it is dominated by the mountains surrounding the irrigation areas. In this region the influence of the complex topography on the ATR overwhelms any signal due to land use itself.

Most urban centers too have a smaller DTR than surrounding rural areas (e.g. Kukla et al., 1986). The DTR deficit in cities is generally between 1–3 K, and, as in irrigated areas, it is larger is summer than in winter (Gallo et al., 1996). The instrument climate record indicates decreasing DTR values in most land cover conditions (Karl et al., 1993), but the decrease is most pronounced in urban areas: the DTR trend for the US historical climatological network (1950–1996) was -0.4 K/century for rural sites and -0.9 K/century for urban sites (Gallo et al., 1999). The overall warming in urban areas during the last century (Linacre and Geerts, 1997, p. 74) was largely due to an increase in daily minima, especially in summer. The daily maximum temperatures changed little. The higher nighttime minima in cities, mainly in summer, also imply a higher mean temperature during the warmest month and hence a larger ATR. The few examples of trends in space and time, given in Section 4, suggest that urbanisation increases mean temperature of the warmest month more than that of the coldest month, and hence increases the ATR in most cases.

In summary, both irrigation and urbanisation tend to decrease the DTR, mainly in summer. The former mainly reduces the daytime maximum while the latter mainly increases the nighttime minimum. As a result, summers are cooler in irrigated areas and warmer in urban areas, implying a smaller ATR in the former and a larger ATR in the latter. The effect of urbanisation on ATR is less uniform and possibly weaker. In some cities, such as Phoenix, the summer maxima and the ATR would probably be higher without the intensive summertime use of water in a desert environment.

This note is merely exploratory. Questions to be answered in further studies include:

- Is a larger ATR in cities, compared to surrounding rural areas, corroborated by a comprehensive station analysis and other observational evidence (e.g. from satellite)?
- Do numerical simulations of the atmospheric circulation and land surface interaction confirm the observed changes in ATR, both in irrigation areas and in cities? And how does the size of an irrigation area or city affect their impact on ATR?
- If so, can these simulated changes be explained through changes in DTR and mean temperature in the warm season, as discussed above?

6. Conclusions

1. The annual range of surface air temperature is reduced by 1–2 K when an arid region such as the Murrumbidgee-Coleambally-Murray Irrigation Area (MIA) is irrigated. The reduced annual range is due to lower temperatures in summer. Also, a ~ 1 K higher dewpoint is found over the MIA in summer, suggesting that the main process explaining the lower annual range is increased evaporation, which reduces daytime maxima in summer.
2. Urbanisation tends to increase the annual range by 0–2 K, mainly through greater urban warming in summer. The higher monthly-mean temperature of the warmest month is mainly due to warmer nights.

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