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Moisture, thermal inertia, and the spatial distributions of near-surface soil and air temperatures: understanding factors that promote microrefugia

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Publication Details

Ashcroft, M. B. & Gollan, J. R. (2013). Moisture, thermal inertia, and the spatial distributions of near-surface soil and air temperatures: understanding factors that promote microrefugia. *Agricultural and Forest Meteorology*, 176 77-89.

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Abstract

Climate change poses significant threats to biodiversity, but some species may be able to escape its effects in small locations with unusual and stable climates (microrefugia). However, there are still great uncertainties about where microrefugia are located, and the exact role that moisture plays in buffering extreme temperatures. In this study we quantified the effects of moisture on the distribution and variability of near-surface soil and air temperatures. We collected hourly 1 cm soil and 5 cm air temperatures and humidities at 111 sites from May 2011 to March 2012. Sites were diverse in terms of elevation (2-1428 m), distance from coast (180 m-403 km), canopy cover (0-100%), topographic exposure, and susceptibility to cold air drainage. We found that variability (diurnal range) of both soil and air temperatures decreased under moister conditions. While air temperatures were related more strongly to humidity, soil temperatures were related more to vapour pressure deficit (VPD). That is, both high temperature and low humidity were required before the VPD was sufficient to dry out the soil and allow soil temperatures to vary. We then used a regional regression approach to model the spatial distribution of minimum and maximum air and soil temperatures for each day over the 10 months in terms of latitude, elevation, canopy cover, distance to coast, cold air drainage potential, and topographic exposure to the south and northwest. We found that elevation was the dominant factor explaining the distribution of soil and air temperatures under moist conditions. Other factors, such as canopy cover and topographic exposure, had a stronger influence on air temperatures whenever humidity was low. However, these factors only affected soil temperatures at times when higher temperatures combined with low humidity to produce higher VPD. Our results provide new insights into how moisture influences the spatial distribution of near-surface soil and air temperatures. Microrefugia will be more apparent under drier conditions, but climate change may affect refugia for soil and air temperatures differently. Higher temperatures will cause VPD to increase more than would be expected by any change in humidity, and refugia in terms of soil temperatures may therefore become increasingly apparent.

Disciplines

Medicine and Health Sciences | Social and Behavioral Sciences

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Abstract

Climate change poses significant threats to biodiversity, but some species may be able to escape its effects in small locations with unusual and stable climates (microrefugia). However, there are still great uncertainties about where microrefugia are located, and the exact role that moisture plays in buffering extreme temperatures. In this study we quantified the effects of moisture on the distribution and variability of near-surface soil and air temperatures. We collected hourly 1cm soil and 5cm air temperatures and humidities at 111 sites from May 2011 to March 2012. Sites were diverse in terms of elevation (2–1428m), distance from coast (180m–403km), canopy cover (0–100%), topographic exposure, and susceptibility to cold air drainage. We found that variability (diurnal range) of both soil and air temperatures decreased under moister conditions. While air temperatures were related more strongly to humidity, soil temperatures were related more to vapour pressure deficit (VPD). That is, both high temperature and low humidity were required before the VPD was sufficient to dry out the soil and allow soil temperatures to vary. We then used a regional regression approach to model the spatial distribution of minimum and maximum air and soil temperatures for each day over the 10 months in terms of latitude, elevation, canopy cover, distance to coast, cold air drainage potential, and topographic exposure to the south and northwest. We found that elevation was the dominant factor explaining the distribution of soil and air temperatures under moist conditions. Other factors, such as canopy cover and topographic exposure, had a stronger influence on air temperatures whenever humidity was low. However, these factors only affected soil temperatures at times when higher temperatures combined with low humidity to produce higher VPD. Our results provide new insights into how moisture influences the spatial distribution of near-surface soil and air temperatures. Microrefugia will be more apparent under drier conditions, but climate change may affect refugia for soil and air temperatures differently. Higher temperatures will cause VPD to increase more than would be expected by any change in humidity, and refugia in terms of soil temperatures may therefore become increasingly apparent.

Keywords: climate change; diurnal range; refugia; soil moisture; temperature variability; topoclimate

1. Introduction

It has been predicted that climate change will cause latitudinal and elevational shifts in species distributions and lead to the possible extinction of 15–37% of species (Hughes, 2000; Thomas et al., 2004). However, these predictions may be overly pessimistic, in part because they are based on macroclimatic conditions and ignore small locations with unusual climates (microrefugia) where species may be able to escape the effects of climate change (Rull, 2009; Ashcroft, 2010; Dobrowski, 2011). Indeed, while ice sheets or extreme aridity may render large regions uninhabitable and force local extinctions or broad-scale migrations, phylogeographic evidence suggests that in unglaciated regions, extinctions and range shifts have been rare and most species have been able to persist through Pleistocene climatic cycles in microrefugia within heterogeneous landscapes (Byrne, 2008; Rull, 2012). Topographic heterogeneity is also recognised as an important buffer against future climate change (Ackerly et al., 2010; Scherrer and Körner, 2011), but we are only just beginning to identify the locations of microrefugia within topographically complex regions and understand how they function (Dobrowski, 2011; Ashcroft et al., 2012; Keppel and Wardell-Johnson, 2012).

There is still a lot of confusion around the exact definition and location of microrefugia (Ashcroft, 2010; Dobrowski, 2011; Keppel and Wardell-Johnson, 2012), and more effort is needed to determine the factors and processes that determine their location (Hampe et al., 2013). An important step towards this goal is determining what causes locations to experience climatic buffering (reduced variability and susceptibility to extremes). Buffering of near-surface temperatures may result from thermal inertia of moist soils (Lu et al., 2009), reduced net radiation fluxes due to canopy cover (Geiger, 1971), or topography that promotes cold air pools or creates shelter from winds (Ashcroft et al., 2009; Dobrowski, 2011; but see also Ashcroft et al. (2012) for evidence that cold air pools actually have higher climatic variability). This buffering of temperatures may cause the near-surface air at these locations to be decoupled from the free atmosphere under some weather conditions (Daly et al., 2010; Pepin et al., 2011), but this decoupling is not present under all weather conditions, and factors such as cloud cover, radiation, rainfall and wind can act across all sites simultaneously and cause some synchronicity in conditions between both coupled and decoupled sites. We argue that climatic buffering needs to be understood not just in terms of decoupling from the atmosphere, but also in terms of the spatially and temporally variable effects of a number of other climate-forcing factors (wind, radiation, soil moisture, cold air drainage potential, canopy cover etc.).

The processes that influence climatic buffering (e.g. radiation, cold air drainage, thermal inertia of soils) generally operate within a few centimetres of the soil surface, and hence localised buffered climates are more likely to be apparent at the soil surface (Chen et al., 1999). However, most meteorological observations are made at a standardised height of 1.5 to 2m, and wind and convection will obscure localised climates at this height. It is difficult to convert standardised observations to near-surface temperatures, as relationships between the two are affected by factors such as cloud cover, time of day, season, wind, canopy cover, topographic exposure and proximity to coast (Wolfe, 1945; Geiger, 1971; Bond-Lamberty et al., 2005; Likso, 2006; Ashcroft and Gollan, 2012). To improve our understanding of climatic buffering and the potential impacts of climate change, it is important that studies focus on near-surface air and soil temperatures rather than standardised observations (Graae et al., 2012; but see also Ashcroft et al., 2009; Holden et al., 2011 for evidence that topoclimatic data may be linked with standardised observations, as this may allow us to produce topoclimatic maps of near surface temperatures in other places or times).

The zone a few centimetres above and below the soil surface is also of primary ecological importance for germinating seeds, sensitive saplings, foraging animals, bushfire fuel moisture and ecological processes such as decomposition, soil respiration and evaporation (Kennedy, 1997; Chen et al., 1999; McVicar et al., 2007; Kustas and Anderson, 2009; Holden and Jolly, 2011; Graae et al., 2012). For example, plants may successfully reproduce only at favourable places and times (their regeneration niche; Ranieri et al., 2012), and the spatial distribution of saplings may be restricted to localised environments rather than the entire species' distribution (McLaughlin and Zavaleta, 2012). If these localised environments provide safe havens that are crucial for the persistence of populations, then accurate, fine-scale climatic data will be needed to understand and predict current and future distributions. As long as the quality of climate data continues to be overlooked as a source of error in species distribution models (Soria-Auza et al., 2010), and insufficient research is undertaken to relate global climate change to ecologically relevant microclimates (Kennedy, 1997), we will be unable to predict future impacts to microclimate-sensitive species (e.g. Roslin et al., 2009).

The magnitude of the differences between surface conditions and standardised observations should not be underestimated. For example, on a hot summer day, standardised observations of 40°C may correspond with ground temperatures of 60°C (Campbell and Norman, 1998; Ashcroft and Gollan, 2012). This introduces a bias that poses a fundamental

problem when we compare, for example, thermal tolerances of species against standardised observations. It has been noted that warming tolerances (differences between a species critical thermal maximum and the mean temperature of warmest quarter at standardised observation height) can be up to 45°C (e.g. Diamond et al., 2012), but a large portion of this is likely due to the differences between standardised observations and near-surface conditions rather than an actual buffer against warming (a portion is also due to the difference between mean temperatures and maximum temperatures). The fact that warming tolerances differ between species in arid and forested environments (e.g. Diamond et al., 2012) is also likely to reflect the different biases between standardised observations and near-surface conditions in these environments, which will obscure any actual differences in their exposure to warming.

Near-surface climatic conditions are also of interest because they are a crucial component of regional and global climate models (RCMs and GCMs), which are the basis of future climate predictions. For example, soil moisture affects the proportion of radiation that results in evaporation instead of temperature rise, results in feedbacks that prolong regional droughts or floods, and forms a crucial component of the planetary boundary layer that drives the underlying climate models (Evans et al., 2011). More research is needed to determine how changes in near-surface soil moisture and climatic conditions will modulate regional climates, and this may also improve our ability to downscale future climate predictions to finer scales (Diffenbaugh et al., 2005; Evans et al., 2011).

The objective of this study was to improve our understanding of how the variability and spatial distribution of near-surface soil and air temperatures are influenced by moisture. The specific goals were: 1) to quantify the variability of soil and air temperatures over short time-scales (diurnal range) and determine the relationships with humidity and vapour pressure deficit (VPD); and, 2) to quantify how the spatial distributions of soil and air temperatures change over time, and explain these temporal trends in terms of how the effects of different climate forcing factors change under different moisture levels. To maximise the generality of results, we made observations over a large region with a diversity of habitats and topographic positions and a broad range of both elevation and distance to coast.

2. Material and methods

2.1 Study area and observation locations

The study area was a large (~500 km by 300 km) region of coastal NSW, Australia that includes the Hunter and Macquarie Valleys (31.0–33.9°S, 147.5–152.6°E; Fig. 1). Much of

the area has been cleared for farming or large open cut mines, but there are also many national parks and state forests. Vegetation communities include coastal swamps and dune communities, open eucalypt forests, temperate and sub-tropical rainforests, upland swamps, woodlands, and grasslands.

We first established a network of 150 DS1923 hygrochron iButton sensors in the Hunter Valley in May 2009, and have recorded hourly temperature and humidity, with four gaps of ~7 days and one gap of 20 days, until present (March 2012; Ashcroft and Gollan, 2012). This was further supplemented by an additional 100 sensors in December 2010, extending the study area further west into the Macquarie Valley, and increasing the density of observations in coastal and high elevation areas (Fig. 1). Our sites covered a diversity of environments in terms of distance to coast (180m to 403km), elevation (2m to 1428m a.m.s.l.), topographic exposure, canopy cover, and land use. We have a complete record (984 days between 1st June 2009 and 25th March 2012) at 94 of the original 150 sites (Fig. 1; we get data from 92.3% of sensors per 3–6 month period on average, with 2.3% disturbed, 2.3% malfunctioned, and the remainder inaccessible, lost, or with unusable data for other reasons).

We buried DS1922L thermochron iButtons 1cm below the soil at 247 of our 250 sites in May 2011, but experienced a higher failure rate (data retrieved from 74.7% of sensors on average, 18.1% malfunctioned, 5.4% lost, 1.8% disturbed or inaccessible) and so only 111 sites have a complete record of both soil and air temperatures between May 27th 2011 and March 25th 2012 (Fig. 1). Most of the analysis in this paper uses only the data from these 111 sites, although we also refer to the 94 sites with a complete three-year record of air data and other individual sites as necessary.

The hygrochron iButtons were suspended ~5cm above the soil surface inside inverted plastic containers with holes for airflow (See Ashcroft and Gollan, 2012 for full details). They were programmed to record at high resolution (precision of 0.0625°C / 0.04%RH; hourly observations for 85 days at a time) for the first year, but this was later deemed unnecessary due to the large differences between sites and they were since configured to record at low resolution (precision of 0.5°C / 0.6%RH; hourly observations for 170 days at a time) so that site visits could be less frequent. Both temperature and humidity observations were software corrected using internal, sensor specific, factory supplied calibration data. The humidity observations needed further adjustment, as the iButtons are capacitive and can saturate under humid conditions producing values higher than 100%. The correction supplied by the manufacturer was deemed unsatisfactory, and we developed our own correction based

on our experience over the ~3 years of observations (see section S1 in supplementary material). We verified corrected observations at each site against the full three-year time series at sites with complete data, as well as with those at nearby sites (section S1).

It should be noted that there was high inter-annual variability in temperatures and precipitation over the three years of observations. The first summer contained a hot, dry period in late November and early December which was associated with bushfires in parts of the study area. In the second summer, there was a high rainfall period at a similar time of year and flooding over much of the study area. The third summer contained high and consistent rainfall over much of the season, and indeed the average of the 94 sites with continuous observations over the three years (Fig. S1 in supplementary material) shows that the peak of the 30-day moving average of maximum daily temperatures was approximately 5°C lower in the third year than the previous two. As the analysis of soil and air temperatures focuses on this third year, results will reflect the cold and wet summer experienced during the recording period.

2.2 Moisture and short-term temperature variability

We quantified spatial and temporal trends in diurnal ranges, using them as a surrogate for climatic buffering (reduced variability and susceptibility to extremes). We assumed that lower diurnal ranges would occur at sites or times where climates were more buffered, and predicted that this this would occur under moister conditions. We tested this by producing linear least squares regression models for soil and air diurnal ranges using predictors of minimum humidity and maximum vapour pressure deficit (VPD). Humidity was included as a direct measure of moisture in the air, and this was compared with VPD because the latter better reflects the drying power of the air (i.e. potential evaporation). Humidity and VPD are related to each other and can be highly collinear under low temperature variability, but they reflect different physical processes and display different seasonal trends (VPD increases with decreasing humidity in summer, but is consistently low in winter when temperatures remain low; Fig. 2). It is important to determine which of the two factors have a stronger effect on climatic buffering as it will give an indication of the underlying processes that buffer sites from extreme climates and lead to a better understanding of how buffering will be affected under future climatic conditions.

For each day and site (304 days $\times$ 111 sites = 33,744 samples) we calculated the minimum and maximum daily humidity and air temperature as well as the minimum and

maximum daily soil temperature. For each of the 33,744 samples we then calculated the maximum daily VPD in Pascals using the formula:

$$\text{VPD} = \text{SVP} * (100 - \text{RH}) / 100$$

where RH was the minimum daily relative humidity of the sample, and SVP was calculated using the formula:

$$\text{SVP} = 610.7 * 10^{7.5T/(237.3+T)}$$

where T was the maximum daily air temperature in degrees Celsius (Monteith and Unsworth, 1990).

We tested for relationships between moisture and diurnal ranges in terms of both temporal patterns (n = 304 days; using average diurnal range, humidity and VPD over 111 sites for each day) and spatial patterns (n = 111 sites; using average diurnal range, minimum humidity and VPD for the days in summer and winter months separately). We also cautiously tested whether the relationships were consistent enough to simultaneously capture both spatial and temporal trends (n = 111 sites × 304 days = 33,744 samples), but we acknowledge that such an approach is more pseudo-replicated (multiple samples at same site or on same day), and it was only used to confirm the generality of the separate spatial and temporal patterns identified above. We also tested whether relationships were robust to reducing pseudoreplication further (only including one sample every ten days to reduce temporal autocorrelation), but this did not change the identified trends (results not shown). Spatial autocorrelation was unlikely to affect results as sites were spread by ~10km on average (Fig. 1), and efforts were made to ensure neighbouring sites were from different environments (Ashcroft and Gollan, 2012).

2.3 Factors affecting the spatial distribution of air and soil temperatures

For each of the 304 days between May 27th 2011 and March 25th 2012 we used a regional (linear least squares) regression approach (Lookingbill and Urban, 2003; Daly, 2006) to relate daily minimum and maximum soil and air temperatures at our 111 sites to a variety of climate-forcing factors. These factors included latitude, elevation, canopy cover (a remotely sensed estimate), relative elevation (elevation above the minimum elevation within 500m - a surrogate for cold air drainage potential), distance to coast, and topographic exposure to the northwest and south. All predictors were 25m resolution grids, and their sources have been described and justified by Ashcroft and Gollan (2012). For example, relative elevation has been shown to be better than alternative predictors of cold air drainage potential based on

flow accumulation or distance to drainage lines (Ashcroft and Gollan, 2012). Predictors were transformed as necessary to ensure linearity: cold air drainage potential was transformed as $\log(\text{relative elevation} + 10)$; canopy cover was not transformed for minimum temperatures, but was transformed using canopy cover^3 for maximum temperatures; and distance to coast was transformed as $\log(\text{distance to coast} + x)$, where x was between 500 and 500,000 m according to what resulted in the highest r^2 in each of the regional regressions. Each of the coefficients in the regressions was converted to an effect size (coefficient $\times$ range of predictor) so that the relative effects of different predictors on spatial patterns could be compared (the coefficients themselves have different units and cannot be compared directly). Interquartile range could also be used if predictor ranges were influenced by outlying observations, but our sites were chosen to avoid outliers. For consistency, all models contained the full seven predictors named above, although we only focus on results for four dominant factors in this study (i.e. elevation, canopy cover, cold air drainage potential and distance to coast for minimum soil and air temperatures; elevation, canopy cover, exposure to northwest and distance to coast for maximum soil and air temperatures). Each of the models (304 models for minimum and maximum air and soil temperatures; one for each day) was based on GIS layers such that they could be used to produce a 25m resolution climate grid across our study region (see Ashcroft & Gollan, 2012).

To improve our understanding of how varying moisture levels affected the spatial distribution of soil and air temperatures, we used our 304 day time-series of spatial models. For each day, we calculated the effect sizes of the four dominant climate-forcing factors in the models for soil and air temperatures; used least squares regressions to see how the effects for soil and air temperatures related to each other; and, used least squares regression to see how the effect sizes were affected by the humidity and VPD on the respective days.

3. Results

3.1 Short-term temperature variability

Average daily minimum air temperatures across our 111 sites varied between ~ 0 – 10°C in winter and ~ 10 – 20°C in summer (Fig. 2). Average daily maximum air temperatures varied between ~ 15 – 25°C in winter and ~ 20 – 40°C in summer. Average soil temperatures were generally warmer in terms of minimums, and colder in terms of maximums, with reduced diurnal range especially noticeable in winter when there was lower VPD (Fig. 2). Soil temperatures also had less day-to-day variability, with consecutive days differing by

1.4°C/0.8°C in terms of average maximum and minimum soil temperatures, while average air temperatures on consecutive days differed by 2.4°C/1.2°C, respectively.

There were strong correlations between maximum daily VPD and minimum daily humidity ($r^2 = 0.69$) and between soil and air diurnal ranges ($r^2 = 0.77$). However, the seasonal trends of VPD and humidity differed, as did the seasonal trends in soil and air diurnal ranges, and it is important to note they are not equivalent (Fig. 2). In terms of temporal trends ($n = 304$ days), we found air diurnal range was more strongly and linearly correlated with humidity ($r^2 = 0.88$ versus $r^2 = 0.76$ for VPD), while soil diurnal range was more strongly and linearly correlated with VPD ($r^2 = 0.87$ versus $r^2 = 0.74$ for humidity; Fig. 3). The results were thus consistent with air temperatures being more variable as the air became drier (lower humidity), but soil temperatures only became more variable when the drying power (VPD) of that air became high enough to dry out the soil (higher VPD is more strongly associated with higher evapotranspiration and requires both high temperature and low humidity; Fig. 2, 3). This difference between VPD and humidity can explain why soil diurnal ranges remain low in winter when VPD is consistently low, but have similar variability to air temperatures in summer when VPD is higher due to higher temperatures (Fig. 2). In contrast, air diurnal ranges can be high throughout the year when there is low humidity.

In terms of the average diurnal range of sites ($n = 111$ sites), air diurnal ranges were more weakly correlated with humidity than VPD, but the relationships for the summer and winter periods were almost identical, while those for VPD were vastly different (Fig. 4). Therefore humidity was a more consistent predictor of air diurnal range. Soil diurnal ranges were once again more strongly correlated with VPD than humidity, although differences were less apparent than for temporal trends. The results for the average diurnal range for sites (Fig. 4) were therefore consistent with the average diurnal range for days (Fig. 3), with similar relationships identified. Spatial and temporal analyses both indicated that humidity was more strongly and consistently correlated with the variability of air temperatures while VPD was more strongly and consistently correlated with the variability of soil temperatures.

While the combined spatial-temporal analysis ($n = 304 \text{ days} \times 111 \text{ sites} = 33,744$) was more pseudo-replicated, it also provided support for these trends. Air diurnal ranges were more strongly correlated with humidity ($r^2 = 0.73$) than VPD ($r^2 = 0.65$) and soil diurnal ranges were more strongly correlated with VPD ($r^2 = 0.63$) than humidity ($r^2 = 0.50$). These results are important because they show that the relationships are consistent across a broad

study area (~500km by 300km), a diverse range of habitats, and over at least 10 months of the year. They also demonstrate that the amount of buffering a site receives will differ according to whether you are looking at air or soil temperatures, and these are affected by different processes and factors (low humidity versus a combination of low humidity and high temperature).

3.2 Factors affecting the spatial distribution of minimum soil and air temperatures

The spatial models for daily minimum air and soil temperatures ($n = 111$ sites) were strong and consistent (average r^2 of 0.69 and 0.71 respectively over the models for the 304 days; s.d. of 0.09 and 0.09). There were also strong correlations between the respective effect sizes (coefficients $\times$ ranges) of the different climate forcing factors in the models for minimum soil and air temperatures ($r^2 > 0.61$; Fig. 5 right column), but the coefficients were not equal. Each of the four factors examined had larger and more variable effects on air temperatures than soil temperatures, with the slopes of the respective relationships ranging from 0.46 to 0.66 (Fig. 5). The larger effect sizes also meant there was more spatial variation in air temperatures (14.2°C) than soil temperatures (11.1°C) when averaged over the 304 days of our study.

The largest difference between the models of air and soil temperatures was apparent in the effect sizes of the canopy cover predictor (Fig. 5, second row). The effect of canopy cover on soil temperatures was much reduced, and indeed in summer it had a warming effect on minimum air temperatures yet a cooling effect on minimum soil temperatures. Cold air drainage potential (relative elevation) also had much less effect on minimum soil temperatures than air temperatures, and also appeared to have a positive effect on air temperatures yet a negative effect on soil temperatures at some times in summer. Coastal effects were more consistent between soil and air temperatures, with both generally agreeing on whether it was cooler or warmer near the coast (Fig. 5). The effects of elevation varied from ~5°C–15°C (lapse rates of 3.5–10.5°C/1000m), with lower lapse rates apparent at cooler and moister times of the year. When lapse rates were low, air lapse rates were lower than those for soil, but when lapse rates were high, the air lapse rates were higher (Fig. 5).

3.3 Factors affecting the spatial distribution of maximum soil and air temperatures

The models for air and soil maximum temperatures were moderately strong and consistent (average r^2 of 0.57 and 0.51 respectively; s.d. of 0.11 and 0.09), and there was similar

average spatial variation across the 304 days (24.4°C and 24.6°C respectively). There were moderate to strong relationships between the respective effect sizes (coefficients $\times$ ranges) of the climate forcing factors in the air and soil models in terms of coastal distance ($r^2 = 0.53$), exposure to the northwest ($r^2 = 0.80$) and canopy cover ($r^2 = 0.60$), but not for elevation ($r^2 = 0.06$; Fig. 6 right column). However, there was a noticeable trend that the effect sizes of all factors other than elevation was diminished in winter and during wet times in summer (Fig. 6 left column). Indeed, the coefficients in the models for soil temperatures were also related to the VPD on the respective day (r^2 of 0.40–0.56; Fig. 7 right column) and multiple regressions that considered both VPD and the respective coefficients for air temperatures boosted the correlations (r^2 of 0.73–0.81). In other words, the coefficients in the models for soil temperatures were related to both the coefficients of the respective models for air temperatures, as well as the average VPD of the respective days. Under cool or moist conditions (low VPD), soil temperatures were largely unaffected by canopy cover, exposure to the northwest and distance to coast, and soil temperatures were determined largely by elevation with a lapse rate of $\sim 5.9^\circ\text{C}/1000\text{m}$ (effect of 8.5°C over elevational range of 1426m; Fig. 7 right column). However, as the VPD increased, the lapse rate increased and canopy cover, exposure to the northwest and coastal distance had progressively more effect on maximum soil temperatures.

The effects of canopy cover, exposure to the northwest and distance to coast also had more effect on maximum air temperatures when humidity was low (Fig. 7 left column). In addition, the effects of cold air drainage and canopy cover on the spatial distribution of minimum air temperatures also increased as humidity became lower (Fig. 8 left column). However, VPD did not affect the influence of these factors on minimum soil temperatures (Fig. 8 right column).

Each of the models for minimum and maximum temperatures can be displayed as a 25m resolution spatial map. We used January 3rd (VPD = 4280, humidity = 40%) and June 8th (VPD = 891, humidity = 52%) as examples to determine where the coldest sites were located (Fig. 9). Under low VPD conditions in winter (June), the map of maximum soil temperatures closely resembled the elevation of the study area (Fig. 1, 9). However, the humidity was quite low on this day, and a stronger coastal effect was apparent in the map of June air temperatures. This coastal effect was even stronger for summer (January) soil temperatures, and stronger again for summer air temperatures (Fig. 9). When we examined finer scale features, it was apparent that sheltered gorges and locations with high canopy cover were also

more likely to be amongst the coldest sites for January maximum temperatures in particular. When it was moist, topography, canopy cover and coastal effects were weak, temperature was well correlated with elevation and refugia were not as apparent. Under drier conditions, topography and other factors had stronger effects and refugia were more apparent because sheltered sites remained moist and were buffered from extreme temperatures. Therefore, refugia will be more apparent in summer than winter and for air temperatures rather than soil temperatures. It is also likely they will be more apparent in drier years, as our study was conducted in a relatively wet summer (Fig. S1).

4. Discussion

Our results provide new insights into how moisture influences the spatial distribution of near-surface soil and air temperatures across a diversity of environments in a large region. Moisture reduces the temporal variability of soil and air temperatures and is likely to play a key role in the processes buffering microrefugia from regional climate change. Under moist conditions, we found that the diurnal range of both soil and air temperatures was reduced (Fig. 3), and elevation had a more dominant effect on the spatial distribution of temperatures (Fig. 7, 8). Under drier conditions, diurnal ranges increased, and cold air drainage, topographic exposure, coastal influences and canopy cover had larger effects, altering the spatial distribution of temperatures and making microrefugia for air temperatures in particular more apparent (Fig. 9). Sites that maintained moister conditions were buffered from temperature fluctuations (Fig. 4), which will help them function as microrefugia (Rull, 2009; Ashcroft et al., 2012).

It is already well appreciated that moisture in the air or soil reduces temperature variability (Bennie et al., 2008; Lu et al., 2009) and may buffer sites from climate change (Fridley, 2009; Ashcroft, 2010). It is also known that anticyclonic conditions (dry, clear skies) increase the effects of factors such as topographic exposure, canopy cover and cold air drainage (Dobrowski, 2011; Jones et al., 2011). However, moisture displays a high degree of spatial and temporal variability (Bond-Lamberty et al., 2005), is difficult to predict, and the novelty of our study is associating moisture with the spatial distribution of near-surface soil and air temperatures across diverse environments, seasons and synoptic conditions. Rainfall reduces spatial variability in soil moisture (Buttafuoco and Castrignandò, 2005) and leads to temperature patterns that are determined largely by elevation (Ashcroft and Gollan, 2012), but predicting moisture distributions under drier conditions remains more problematic as they

are influenced by a wider variety of factors (Lookingbill and Urban, 2004). Microrefugia will be more apparent under these drier conditions, as the buffering offered by canopy cover, topography and coastal influences will be more apparent when exposed sites have dried out but sheltered sites remain moist.

Our results highlight that moisture affects near-surface air and soil temperatures in different ways. Air temperatures have high variability whenever humidity is low, while both high temperature and low humidity (high VPD) are required before soils dry out and soil temperatures become variable. In winter, VPD remains low, and soil temperatures consistently display low diurnal range, even in our temperate study area where there is no persistent snowpack (Fig. 2; see also Bond-Lamberty et al., 2005). The low winter VPD also reduces the effects of canopy cover, topographic exposure and coastal influences on maximum soil temperatures (Fig. 7; see also Breshears et al., 1998), even though these factors can still have a large influence on the distribution of air temperatures (Fig. 6). However, soil moisture is not just determined by VPD, with soil texture, leaf litter and topographic position also having an effect (Buttafuoco and Castrignanò, 2005). Sites that can maintain dry conditions throughout the winter months (e.g. by good drainage, high solar exposure, sandy soils) can also maintain high temperatures, and some of our sites experienced maximum temperatures 15°C higher than average during the winter months (e.g. Fig. S6 in supplementary material). Sites such as these may be important for maintaining populations of ectothermic animals, such as insects, over the winter months. Matching biotic data with fine-scale data such as ours would provide much needed insight into the role of variability as a driver of community dynamics. As noted by Adler et al., (2006), the ecological impacts of increased climate variability are poorly understood, especially in comparison with those posed by increasing mean temperatures.

One shortcoming of our study is that it only examines two heights—1cm below the soil and 5cm above. Previous studies have shown that diurnal ranges decrease deeper into the soil and further off the ground (Geiger, 1971; Campbell and Norman, 1998; Bond-Lamberty et al., 2005), and our near-surface observations are likely to capture the higher spatial and temporal variability of temperatures that occurs near the surface (Campbell and Norman, 1998; Chen et al., 1999). As discussed earlier, near-surface temperatures are important for many ecological processes (Graae et al., 2012), and should be considered more in ecological research (Kennedy, 1997). However, our results demonstrate the complexities and differences in converting from one observation height to another, and the trends identified here will only

reflect conditions at the height/depth we made observations. Indeed, soil moisture will vary with depth, and shallower soil depths may be able to dry out in direct sunlight in winter even if the VPD is low, while soils at deeper depths may remain moist longer than predicted based on 1cm soil temperatures.

There are undoubtedly variations in climate within the 25m cells used in our study, despite the fact our sensors were placed in relatively homogenous habitats. The results of our study are meant to capture the general topoclimatic trend across the landscape better than macroclimate, but without considering microclimatic variation (Ashcroft and Gollan, 2012). The grids are not designed to capture the microclimate of species that can exist in small areas within these cells, or within specific microclimates such as under rocks or inside tree-hollows. They will be most useful for explaining complex landscape scale vegetation mosaics, such as patches of rainforest amongst a matrix of woodlands and eucalypt forests (Ashcroft et al., 2008).

Keppel and Wardell-Johnson (2012) recently hypothesised that there may be tipping points in the buffering capacity of microrefugia, with reductions in rainfall possibly causing a loss of resilience. The results of our study supported this hypothesis. Diurnal ranges, and therefore exposure to extreme temperatures, increased when it was drier (Fig. 3, 4). A number of our partially sheltered or woodland sites were able to maintain moist conditions and avoid extreme temperatures in the wetter third year of our data collection, but exceeded a tipping point and were unable to remain moist in the drier years. While sites were approximately 6°C warmer on average in the drier first year than the wetter third year, there were eleven sites which were 9–16°C warmer because they lost their microclimatic buffering (e.g. Fig. S7 in supplementary material). Not all sites exceeded this tipping point, and more sheltered rainforest sites were able to remain buffered in all years. However, even these microrefugia (Ashcroft et al., 2012) may be threatened by further reductions in rainfall in the future. Alternatively, if rainfall increases in the future then surface temperatures may be more buffered and experience more moderate extremes—even if mean air temperatures are increasing.

We found that buffering of air temperatures was associated with humidity whereas soil temperatures were associated with VPD. The implication is that soil and air temperatures will respond differently to future climate change, especially if higher temperatures cause VPD to increase more than would be predicted by any change in humidity. Similar divergent trends in soil and air temperatures have also been predicted with changes in snow depth, a

factor not considered here, and both experimental results and future projections predict a trend of decreasing soil temperatures even though air temperatures are increasing (Groffman et al., 2001; Decker et al., 2003; Brown and DeGaetano, 2011). Sites that may experience the greatest resilience to climate change are those that remain moister, such as sheltered gorges (e.g. Holec and Wild, 2011), forests, or coastal and high elevation sites, even if they are not necessarily the places that experience the coolest temperatures. These sites have the ability to absorb radiation with less change in temperatures, leading to greater resilience and stability in the face of climate change (Lloret et al., 2012; Norris et al., 2012).

Acknowledgements

This research was funded by the NSW Environmental Trust as part of the Great Eastern Ranges Conservation initiative (Grant GER-08-AM01), and as part of Australian Research Council Linkage Project LP100200080 in collaboration with the Australian Museum, University of Technology Sydney, Central West Catchment Management Authority, NSW Office of Water, and the Australian Wetlands and Rivers Centre at the University of New South Wales. We are grateful to more than 200 landowners who granted us permission to access their land, to Dan Ramp for fruitful discussions, and to Andrew Letten for fieldwork assistance.

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Fig. 1 The 250 locations where hourly temperature and humidity were at least partially recorded between June 1st 2009 and March 25th 2012 (a), along with the 94 locations where we have a complete record (b) and the 111 sites where we have both soil and air data from May 27th 2011 to March 25th 2012 (c). The map at the bottom right shows the location of the study area within Australia (147.5–152.6°E, 31.0–33.9°S).

Fig. 2 The average daily minimum and maximum temperature and humidity from 111 sites where we had a complete record of hourly data between May 27th 2011 and March 25th 2012. Air temperatures at 5cm are shown with dashed lines, soil temperatures at 1cm are shown with solid lines. Maximum daily VPD was calculated using maximum daily temperature and minimum daily humidity. Diurnal range was calculated as the difference between maximum and minimum temperatures for both air (dashed line) and soil (solid line).

Fig. 3 For each of the 304 days between May 27th 2011 and March 25th 2012 we calculated the average minimum daily humidity, average maximum daily VPD, and average diurnal range of soil and air temperatures across our 111 sites. The diurnal range of air temperatures was more strongly and linearly related to humidity, while the diurnal range of soil temperatures was more linearly and strongly related to VPD.

Fig. 4 For each of the 111 sites we calculated the average diurnal range of soil and air temperatures for summer (December–February; hollow squares) and winter (June–August; solid circles), and the respective average minimum humidities and maximum VPD. Relationships were similar to those produced using the data for different days (Fig. 3).

Fig. 5 For each of the 304 days of the study, we regressed daily minimum soil and air temperatures at each of 111 sites against latitude, elevation, canopy cover, cold air drainage potential (relative elevation), distance to coast, and exposure to the south and northwest. The effect sizes (coefficients $\times$ range) for four main factors on each day are shown on the left, with the relationships between soil and air models shown on the right.

Fig. 6 For each of the 304 days of the study, we regressed daily maximum soil and air temperatures at each of 111 sites against latitude, elevation, canopy cover, cold air drainage potential (relative elevation), distance to coast, and exposure to the south and northwest. The effect sizes (coefficients $\times$ range) for four main factors on each day are shown on the left, with the relationships between soil and air models shown on the right.

Fig. 7 The effect size of different factors in the models for maximum air temperatures were weakly to moderately correlated with minimum daily humidity (left panels), while the effects on soil temperatures were strongly related to the VPD (right panels). As VPD approached zero, the effects of canopy cover, coastal distance and exposure to the northwest were reduced substantially, and the effect of elevation approached 8.5°C (lapse rate of $5.9^{\circ}\text{C}/1000\text{m}$ over elevational range of 1426m).

Fig. 8 The effect size of different factors in the models for minimum air temperatures were weakly to moderately correlated with minimum daily humidity (left panels), while the soil temperatures were weakly related to the VPD (right panels). With the exception of the effects of cold air drainage and canopy cover on minimum air temperatures, effects of moisture were weaker than the effects on maximum temperatures (Fig. 7).

Fig. 9 The modelled spatial distribution of daily maximum soil and air temperatures for January 3rd 2012 and June 8th 2011. So that different patterns can be compared, we have converted the raw temperature grids to display the percentile of sites that have colder temperatures than each location. Coldest locations are in black, with warmer locations in white. Coldest locations are generally high elevation sites (Fig. 1), coastal sites, and at finer scales, topographically sheltered sites and those with high canopy cover.